VOLUME LV NUMBER 2

THE JOURNAL OF GEOLOGY

March 1947

DRAINAGE DIVERSION IN THE NORTHERN ROCKY MOUNTAINS OF EAST-CENTRAL IDAHO[1]

ALFRED L. ANDERSON
Cornell University

ABSTRACT

The upper Salmon River, many of its tributaries, and other streams draining the northern Rocky Mountains of east-central Idaho flow for many miles in a northeasterly direction and then abruptly change their courses. The Salmon River makes an elbow turn and virtually reverses its direction of flow to the west side of the state, joined below by important barbed tributaries and above by consequent streams which occupy northwestward-trending structural basins. Other northeast-flowing streams enter structural basins and are directed southeast to the Snake River Plain.

Headwater trends and aligned wind gaps across block mountains are indicative of an early drainage system that extended far to the northeast, probably to the Missouri River. That piracy and large-scale faulting have had an important role in disrupting this early drainage is indicated by barbed tributaries, elbows of capture, wind gaps on tilted block mountains, and the presence of consequent streams in fault-made valleys.

Diversion of Salmon River to the west side of the state apparently resulted from capture by headward erosion of a vigorous stream from the west as a consequence of crustal disturbances during late Tertiary and early Quaternary time. Diversion of other streams into northwest-southeast structural basins between block-faulted mountains probably took place at about the same time. By reason of these diversions the Continental Divide has been shifted about 100 miles east of its location in late Tertiary time.

INTRODUCTION

The drainage in the northern Rocky Mountains of east-central Idaho has a peculiar pattern indicative of a complicated and an eventful development. The Salmon River and some of its important tributaries trend across the area in a northeasterly direction almost to the Montana line and Continental Divide where the main river then makes a sharp elbow turn and virtually reverses its direction of flow to cross to the west side of the state, joined in its westward course by tributaries which are barbed and which, until they enter the Salmon, are parallel to the headwater streams (Fig. 1). Other streams, like Big Lost River, also trend in a northeasterly direction, but after some miles they abruptly change their courses and flow as consequent streams in long, intermontane basins either northwest to join the Salmon River or southeast to disappear on reaching the Snake River Plain. Wind gaps aligned with the headwater streams and extending across the mountains bordering the basins indicate that the drainage formerly extended much farther to the northeast than it does at present.

The thesis here presented is that the streams of northeast trend represent the headwaters of a previously established

[1] Presented at the Chicago meeting of the Geological Society of America, December, 1946.

drainage system that extended east of the Continental Divide, perhaps to the Missouri River, and that the existing peculiar stream pattern is the consequence of piracy and large-scale faulting that disrupted the earlier drainage and caused the Continental Divide to be shifted about 100 miles east of its former location.

TOPOGRAPHIC SETTING

As constituted locally, the northern Rocky Mountains are composed in part of dissected uplands and in part of block mountains and intermontane basins like those in Utah and Nevada (Pl. I).

Dissected uplands.—The dissected uplands make up about two-thirds of the area under discussion, composing the

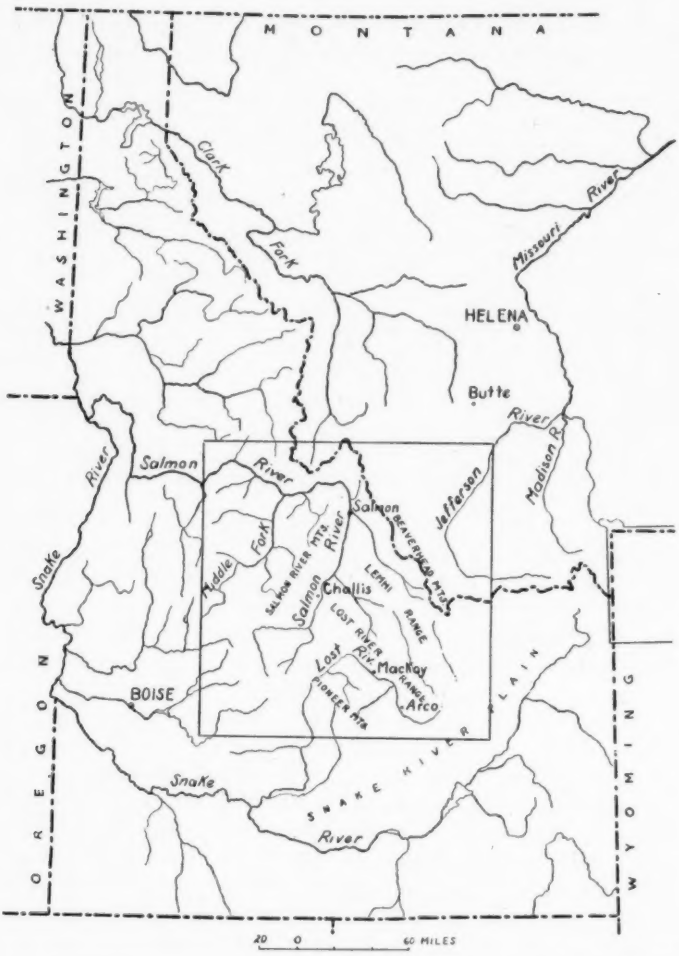
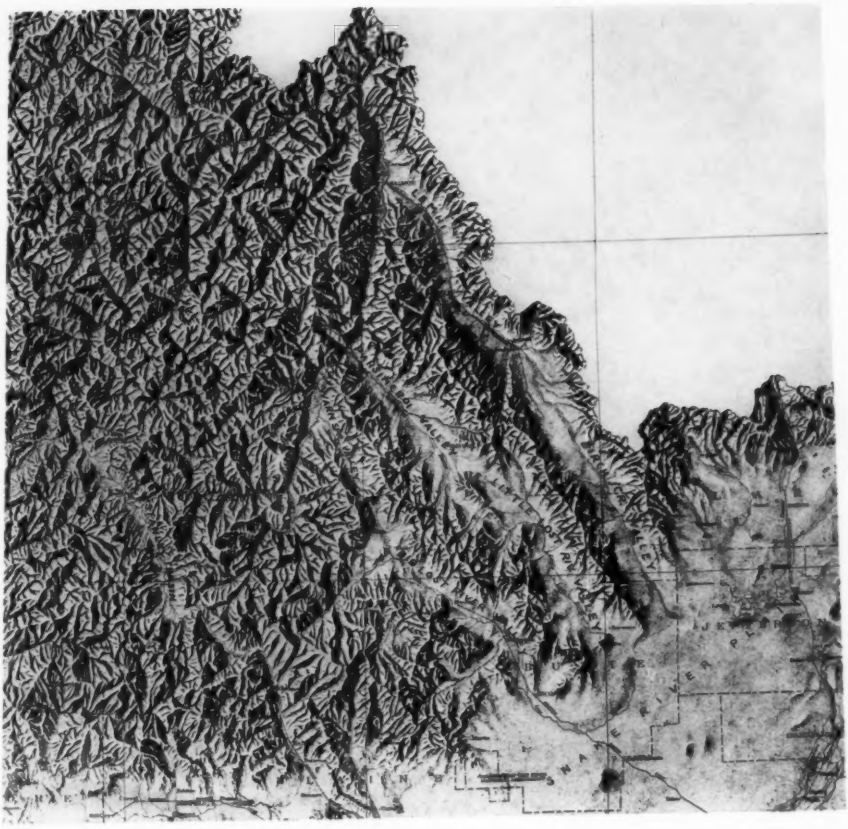


FIG. 1.—Index map showing location of diverted drainage systems in east-central Idaho

PLATE I



Relief map of east-central Idaho showing the northern Rocky Mountains composed locally of dissected uplands and isolated basins and ranges.

PLATE II

A

B

A, Dissected uplands comprising the Salmon River Mountains in the vicinity of Panther Creek. *B*, Desert basin of Birch Creek with the tilted back slope of Lemhi Range in the background.

mountains in the northern and western part. Although the uplands form essentially a continuous mountain group, they have been divided into poorly delimited units which have been named from the drainage systems that dissect them or from prominent physiographic or other features that locally may distinguish them. These units include: the Sawtooth Mountains at the head of Salmon River; the Salmon River Mountains, north and west of Salmon River, mountains which extend from the headwaters of the river in the Sawtooth Mountains to its elbow turn near the Montana line; the White Cloud Peaks, which are near the west edge of an unnamed group between the Sawtooth Mountains and the eastern border of the upland group; the Boulder Mountains southeast of the Sawtooth Mountains and White Cloud Peaks; the Pioneer Mountains between the Boulder Mountains and the Snake River Plain; and the White Knob Mountains between the Pioneer Mountains and the eastern border of the uplands (Fig. 2).

As pointed out by J. B. Umpleby,[2] the dissected uplands composing the largest unit, the Salmon River Mountains, show a marked accordance of ridge levels and bear the remnants of an old erosion surface that has been deeply and intricately trenched by the present drainage system (Pl. II, *A*). This erosion surface may also be recognized in the other mountain units, though not everywhere so well preserved as in the Salmon River Mountains. Although essentially flat over broad areas, the surface shows the effect of differential uplift and locally may form the flanks and buried floors of minor intermontane basins and the crests and back slopes of none-too-prominent mountain blocks. Over much of the Salmon River Mountains the upland surface stands 8,000 to 10,000 feet above sea-level. In some of the other units, ridges rise above 10,000 feet and some peaks above 11,000 feet—the highest, Hyndman Peak in the Pioneer Mountains, reaching 12,078 feet. Despite some topographic discontinuities, however, the upland areas retain all the essential characteristics of a greatly dissected peneplain or erosion surface which has been incised to depths of as much as 5,000 feet by some of the major streams.

Basin and range area.—The basin-and-range part constitutes the southeast third of the area, though actually this part extends well up the eastern border, almost to the elbow turn of Salmon River (Pl. I). It is separated from the upland group by a great intermontane trench that reaches from the Snake River Plain northwest to the Salmon River, a distance of 90 miles, and from that point on by the Salmon River. In addition to this bordering trench, the area embraces three parallel mountain ranges, two of them 90 miles long, the third about 145 miles long, and two intervening trenches 100 and 110 miles long, all extending northwest from the Snake River Plain to the Salmon River (Pl. I). The ranges are known, respectively, from southwest to northeast as the "Lost River Range," the "Lemhi Range," and the "Beaverhead Mountains," the latter forming a part of the Bitterroot Range and carrying on its crest the Continental Divide. The trenches between the ranges and between the Lost River Range and the dissected uplands have been named from the streams that drain them; and, since drainage within them is both northwest and southeast from low, about centrally located valley-floor divides, a part of each trench is known by one name, the remainder, by another. These parts are

[2] "Geology and Ore Deposits of Lemhi County, Idaho," *U.S. Geol. Surv., Bull. 528* (1913), pp. 22–24.

basins of the desert type (Pl. II, *B*), but instead of being designated as basins they are called "valleys," and those separating the Lost River Range from the dissected uplands are referred to as the Warm Spring and Big Lost River valleys, those between the Lost River and Lemhi ranges as the Pahsimeroi and Little Lost River valleys, and those between the Lemhi Range and the Beaverhead Mountains as the Lemhi and Birch Creek valleys (Pl. I). Warm Spring, Pahsimeroi, and Lemhi valleys drain northwest to the Salmon River; the others, to the Snake River Plain.

Topographically the mountain ranges are much alike. Each of them presents an abrupt scarplike front to the broad basins along their southwest borders and less evenly faced slopes on their north-

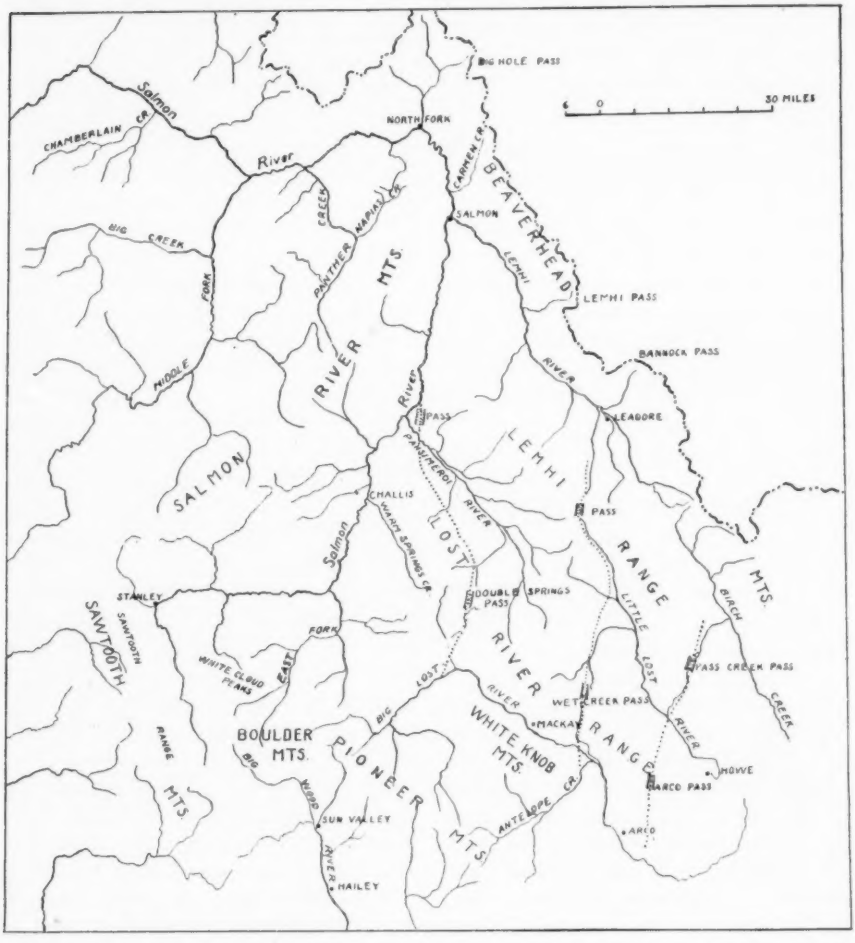


FIG. 2.—Sketch map showing location of mountains, streams, and wind gaps (passes) and former drainage courses (*dotted*).

east sides. In each the crest is close to the southwest border, and each therefore resembles an asymmetrical, steeply tilted block—the southwest side representing the front of the block, the northeast side, the back (Fig. 3). On the southwest sides the regularity of the scarplike faces has been interrupted somewhat by sharply incised V-shaped canyons ending at the bases of the ranges and opening upon extensive alluvial fans (Pl. III, *A*). On the back slopes of both the Lost River and the Lemhi ranges are tilted remnants of an erosion surface like that which distinguishes the upland area. Near the north end of the Lemhi Range the surface becomes broader and flatter and matches the one in the Salmon River Mountains across the Salmon River. The summit erosion surface is exceptionally well preserved in parts of the Beaverhead Mountains, in places being notably flat or gently undulating and little dissected.[3] Across each of the ranges are relatively low natural passes (wind gaps) to which reference will be made later. Except for these passes, the ranges are little notched and present bold, precipitous fronts reminiscent of the Wasatch Range in Utah.

These ranges are among the highest in the state, all rising above 11,000 feet and two of them above 12,000 feet. Lost River Range carries the highest point in the state, Borah Peak at 12,655 feet, which towers 6,000 feet above the bordering basins (Fig. 3). When connected, the high points of these ranges are along an axial line that is parallel to and about 50 miles from the edge of the Snake River Plain. This axis, one of upwarp, parallels the Snake River Plain across the state, passing through Hyndman Peak in the Pioneer Mountains and through other high mountains to the west and northwest. On either side of the line the mountains descend gradually to lower levels, those on the south side to the level of the Snake River Plain.

The intermontane trenches are singularly straight and remarkably uniform in width (Pl. I and Pl. II, *B*). Their northeast borders are especially straight and regular, their southwest borders, where they touch against the back slope of the ranges, being less so. The trenches containing the Birch Creek and Lemhi basins and the Little Lost River and Pahsimeroi basins average 5 to 8 miles wide, although at its head the Pahsimeroi basin is but 2 miles wide and in its lower part as much as 10 miles wide. The trench containing Big Lost River and Warm Spring basins is narrower, mostly 3 to 5 miles wide. All are surfaced by great alluvial slopes extending out from the base of the mountain fronts. As indicated earlier, these trenches drain in opposite directions from low invalley divides. The divide which separates the drainage of Birch Creek from that of the Lemhi River is almost imperceptible, and for about 8 miles has no surface drainage at all. At the divide the trench is as wide as at any place along its course. From about 7,200 feet the Birch Creek basin drops to 5,800 feet where it merges with the Snake River Plain 40 miles to the southeast, and the Lemhi basin descends to 4,000 feet at the Salmon River 70 miles in the opposite direction. The divide between Little Lost River basin and the Pahsimeroi, although restricted by the Donkey Hills, is still several miles wide and is about as imperceptible as the divide between the

[3] P. J. Shenon, "Geology and Ore Deposits of the Birch Creek District, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 27* (1928), pp. 3–4; A. L. Anderson and W. R. Wagner, "Lead-Zinc-Copper Deposits of the Birch Creek District, Clark and Lemhi Counties, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 70* (1944), p. 4.

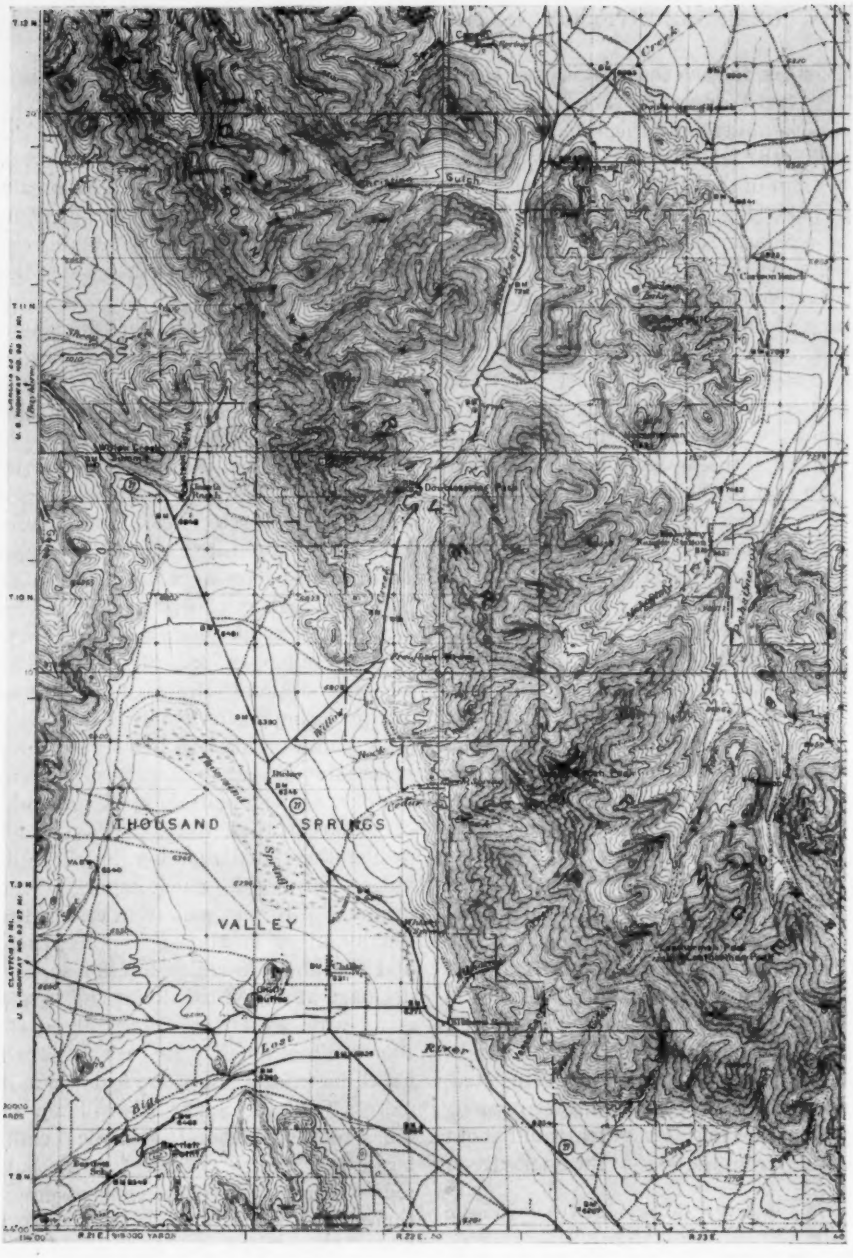


FIG. 3.—Topographic map (Borah Peak quadrangle) showing a part of the asymmetrically block-tilted Lost River Range, its southwestward-facing frontal escarpment, and the wind gap (Double Springs Pass) that crosses the range in alignment with the upper course of Big Lost River.

PLATE III

A

B

A, Lost River Range escarpment from upper Warm Spring Valley, showing aligned faceted spurs, sharply incised V-shaped canyons and fronting alluvial fans. *B*, Double Springs Pass (wind gap) which crosses Lost River Range opposite the place at which Big Lost River enters and flows southeast along Big Lost River Valley. The wind gap is interpreted as a former valley of Big Lost River before its defeat and diversion by block faulting.

PLATE IV

A

B

A, Wet Creek Pass (wind gap) which crosses Lost River in line with Antelope Creek is interpreted as a former course of Antelope Creek before its diversion by block faulting. *B*, Arco Pass (wind gap) near the southeast end of Lost River Range as viewed from Little Lost River Valley. The head of the former Arco River, which occupied the wind gap, has been carried below by the Snake River downwarp and covered by lava.

Birch Creek and Lemhi River basins. From the divide at 6,728 feet above sea-level the Little Lost River basin extends 55 miles southeast to the Snake River Plain and the Pahsimeroi River basin 45 miles northwest to the Salmon River. The divide between the Big Lost River and Warm Spring Creek basins is more pronounced, a prong from the upland area extending across to the base of the Lost River Range (Fig. 3). However, the divide at Willow Creek summit is 7,161 feet and therefore no higher than the one in the Birch Creek–Lemhi River trench. From this divide the Warm Spring Creek drainage extends to the Salmon River 30 miles away and the Big Lost River drainage to the Snake River Plain 60 miles to the southeast. Like the high points on the ranges, the basin divides are close to an axial line that parallels the margin of the Snake River Plain and that coincides with that of the upbowed ranges.

PREVIOUS INTERPRETATIONS

The drainage pattern within the uplands has not previously attracted attention, but considerable interest has been focused on the drainage in the eastern basin-and-range area. Umpleby[4] was first to comment on the origin of the basins and the drainage in them. The basins he regarded as ancient valleys carved in an early erosion surface (peneplain) and the streams that carved them as belonging to a southeast drainage system that extended to the Snake River Plain. The long valley trench occupied by Lemhi River and Birch Creek he regarded as occupied by an ancient stream that headed north of the town of Salmon whose headwaters were later captured by the Salmon River and the divide pushed back to its present location by headward erosion of Lemhi River. He later[5] mentioned re-entrant valleys in the Lost River Range and explained them as belonging to the older drainage system and stated that the one near Arco was once probably occupied by Little Lost River until blocked by lavas in the vicinity of Arco Pass, when the river was diverted into its present channel. The valley of Big Lost River he thought could be accounted for, more certainly than any of the other large valleys, by the erosional activity of the stream that now occupies it, though faulting may have been a factor in the development of the valley, as indicated by the triangular faces on the eastern side.[6]

After making a particular study of the drainage in western Montana and eastern Idaho, W. W. Atwood[7] came to the same conclusion as Umpleby concerning the southeast drainage, but he stressed warping and faulting in the development of the valleys (basins).

O. E. Meinzer[8] later called attention to the contrast between the straight, well-defined valley front on the northeast side of the Pahsimeroi Valley and the indefinite border on the southwest side and suggested faulting as the chief means of valley depression.

V. R. D. Kirkham[9] then expressed a

[4] P. 30 of ftn. 2.

[5] "Geology and Ore Deposits of the Mackay Region, Idaho," *U.S. Geol. Surv., Prof. Paper 97* (1917), pp. 19–21.

[6] Pp. 20–21 of ftn. 5.

[7] "The Physiographic Conditions at Butte, Montana, and Bingham, Utah, When the Copper Ores of These Districts Were Enriched," *Econ. Geol.*, Vol. XI (1916), pp. 706–21.

[8] "Ground Water in Pahsimeroi Valley, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 9* (1924), p. 16.

[9] "A Geologic Reconnaissance of Clark and Jefferson and Parts of Butte, Custer, Fremont, Lemhi, and Madison Counties, Idaho," *Idaho Bur. Mines and Geol. Pamph. No. 19* (1927), p. 11.

view that the depressions were undoubtedly structural and antecedent to the streams that now occupy them but that the relations of several re-entrant valleys, which have a decided northeast trend, indicated that the ancient drainage flowed northward and not to the south as proposed by Umpleby and Atwood.

Soon after, Shenon[10] pointed out that block faulting, in places with a vertical displacement of 3,500 feet, defines the wide valley in which Birch Creek flows and that the relatively straight, clean-cut valley wall with well-defined faceted spurs extends into the Lemhi basin, indicating extensive development of the faulting. Having been over much of the area in question and after a careful review of the literature, he accepted the view of Umpleby and Atwood for a southeastward-flowing drainage system, despite the contradictory evidence pointed out by Kirkham, which, he stated, the broader evidence apparently failed to support.

C. P. Ross[11] has since suggested an origin for the Warm Spring Creek basin and the northwest end of the Lost River Range other than by faulting, but E. M. Baldwin,[12] working in the same region, has concluded that Lost River Range is a tilted fault block and that the bordering basins are fault-controlled.

More recently, the writer[13] has confirmed the fault-block origin of Birch Creek Valley and has further evidence to back up an earlier publication[14] that the three ranges have been produced by block faulting and that the "valleys," so called, are structural basins. In neither of these reports were the drainage problems discussed. Evidence is now to be presented to show that the former drainage was northeast rather than southeast and that capture of Salmon River and block faulting have succeeded in beheading the drainage and directing it into its present channels.

EVIDENCE OF EARLY NORTHEAST DRAINAGE

Evidence of an early northeast drainage system is preserved in the present headwater streams, in the northeast alignment of wind gaps across the block ranges as prolongations of these streams, and in the distribution pattern of Tertiary volcanics.

Headwater drainage.—The evidence afforded by headwater streams is to be found largely in the dissected upland area where the drainage pattern has been least modified by differential crustal movements. The upper Salmon River itself offers a most striking clue to the direction of the early drainage (Fig. 2). For the first 40 miles of its course it flows in a northwesterly direction along virtually the full length of Stanley Basin, an intermontane fault-made basin extending along the base of the Sawtooth Range, a tilted-block segment of the Sawtooth Mountains (Pl. I). Near the settlement of Stanley it leaves the wide alluvial-floored depression and flows eastward in a deep, steep-sided valley that widens somewhat downstream. Near its junction with its East Fork

[10] P. 4 of ftn. 3.

[11] "Erosion in the Lost River Range, Idaho," abstract in *Jour. Wash. Acad. Sci.*, Vol. XXVIII (1938), p. 415.

[12] "Structure and Stratigraphy of the Northern Half of Lost River Range, Idaho" (Ph.D. dissertation, Cornell University, 1943).

[13] P. 13 of ftn. 3.

[14] A. L. Anderson, "A Preliminary Report on Recent Block Faulting in Idaho," *Northwest Science*, Vol. VIII (1934), pp. 17–28.

(Fig. 2) it changes its course from east to northeast, and its valley continues to widen. Just above the town of Challis it debouches into the comparatively broad expanse of Warm Spring Creek basin (Pl. I), skirts the lower end of the basin, and then a short distance below Challis enters the deep narrow canyon that separates the Lost River Range from the Salmon River Mountains. Emerging from the canyon, the river, which continues to flow in a north-northeasterly direction, skirts the lower end of the broad Pahsimeroi River basin and then enters the deep 30-mile-long canyon that cuts across the upland area and separates the Lemhi Range from the Salmon River Mountains. About 10 miles above (south of) the town of Salmon the river emerges from the canyon into the broad basin of Lemhi River and flows in a more northerly direction along the lower northwest edge of the basin to a point about 10 miles below (north of) Salmon, where it enters a canyon from which it never again emerges. As it enters this canyon, it swings from north to northwest and then at the settlement of North Fork makes a right-angle turn and from there crosses to the west side of the state.

The main tributaries of the upper Salmon also conform with it in trend. East Fork, which rises in the White Cloud Peaks and Boulder Mountains area, flows in a northeasterly direction aligned with the Salmon River at and below the point about 35 miles below Stanley where the Salmon changes its course from east to northeast (Fig. 2). The Middle Fork of Salmon River, which has its head at or near the northwest end of the Sawtooth Range (Fig. 2), flows in a northeasterly direction to its junction with the main Salmon about 35 miles below North Fork. Its course is almost exactly parallel to that of East Fork and Salmon River below the East Fork junction. Its principal tributary, Big Creek, joins from the west and from its head to its mouth flows nearly due east. Chamberlain Creek is another tributary of the main Salmon that flows in an east-northeasterly direction. It joins the main Salmon a few miles below the mouth of Middle Fork (Fig. 2). Panther Creek is another important tributary, which, except for its lower few miles, flows in a northeasterly direction. At the point where its course changes from northeast to northwest it is joined by Napias Creek, which, though flowing southwest, is aligned with the upper course of Panther Creek.

The northeast trend of the headwater drainage is also shown by streams other than the Salmon and its main tributaries. Big Lost River, which rises in the Boulder and Pioneer Mountains, flows northeast until it reaches the broad floor of Big Lost River Valley (Pl. I and Fig. 2) where it turns sharply southeast and follows the front of Lost River Range to the Snake River Plain (Pl. I). Antelope Creek, a tributary of Big Lost River, rises in the Pioneer Mountains and flows northeast to its junction with Big Lost River in Big Lost River Valley. Even the head of Pahsimeroi River, which has its source in the Lost River Range, flows obliquely across the back slope of the range (Fig. 3) in a general north-northeasterly direction until it emerges upon the broad floor of Pahsimeroi Valley (basin). Other streams that head on the northeast sides of both the Lost River and Lemhi ranges also trend obliquely across instead of directly down the back slopes, thus conforming with the northeast drainage trends. As they debouch upon the broad plains at the foot of the ranges, they are directed

either northwest or southeast along the basins.

Wind gaps.—The existence of wind gaps or passes, which are aligned in a northeast-southwest direction across the block ranges, is also evidence of an earlier northeast drainage. One of the most impressive of these wind gaps is Double Springs Pass (Fig. 3) which extends across Lost River Range in a north-northeasterly direction opposite the point at which Big Lost River emerges from the uplands into Big Lost River Valley (basin). This pass (Pl. III, *B*) is cut 3,000 to 4,000 feet into the range, its high point of 8,318 feet being in line with the crest of the range and close to its southwest border. The pass constitutes a transverse valley tilted and flaring toward the Pahsimeroi Valley basin. The old valley course is partly choked with morainal debris and outwash from the bordering slopes, and earlier stream gravels are not exposed. The valley is not occupied by permanent streams, and much of the drainage today is subsurface. This old valley crossing is in direct line with Big Lost River as it emerges from the upland area. In crossing the range the old valley swings to the north-northeast and shows alignment with a wind gap that cuts across a corner of Lemhi Range very close to the lower end of Pahsimeroi Valley (Fig. 2). This second wind gap is about 1,000 feet above the Pahsimeroi and Salmon River Valley floors and appears as a deep notch connecting the Pahsimeroi Valley basin with the Snake River Canyon. If Big Lost River were extended through Double Springs Pass and this second wind gap, it would make a perfectly normal tributary of the Salmon River.

Wet Creek Pass (Fig. 2 and Pl. IV, *A*) is another wind gap that forms a deep notch across Lost River Range, in this case essentially opposite the mouth of Antelope Creek. It has been described by Kirkham[15] as a re-entrant valley and has been used by him as evidence of an old northeastward drainage course. This wind gap repeats the essential features of Double Springs Pass. It cuts deeply into the range, its floor being about 7,500 feet above sea-level, and forms a transverse valley tilted northeast. It is floored with gravel washed down from bordering slopes, but gravel caps on ridges carved by Wet Creek in the floor of the valley may consist of old stream gravels. This valley trends north-northeast and is aligned with Antelope Creek as Antelope Creek approaches the edge of the upland area. It is also in line with a long and deep wind gap that extends across the Lemhi Range at the head of Little Lost River (Fig. 2). This second wind gap may be seen to good advantage from the Lemhi Valley near the town of Leadore. Whether Bannock Pass in the Beaverhead Mountains is a prolongation of these wind gaps has not been determined.

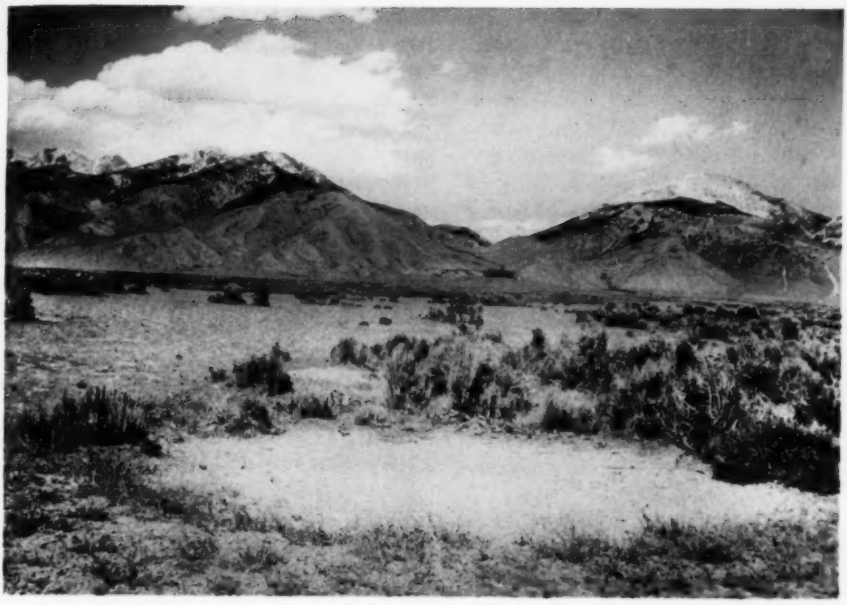
Arco Pass (Fig. 2 and Pl. IV, *B*) close to the southeast end of Lost River Range also attracted the attention of both Umpleby[16] and Kirkham[17] and prompted each to regard it as a former drainage course—regarded by Umpleby as a southwest drainage course, by Kirkham as a northeast drainage course. This pass, or "re-entrant valley" as called by Kirkham, extends in a north-northeasterly direction; but, unlike the other passes along the Lost River Range, it is not opposite any present streams. Instead, it opens upon the Snake River Plain, standing about 1,500 feet above the surface of the plain. The pass is fairly broad

[15] Pp. 10–11 of ftn. 9.

[16] P. 19 of ftn. 5.

[17] Pp. 10–11 of ftn. 9.

PLATE V



Pass Creek Pass (wind gap) which extends across Lemhi Range in line with Arco Pass and was formerly occupied by the defeated and no longer existent Arco River.

(Pl. IV, *B*); and, like the other passes, its floor is tilted northeast and merges with the broad expanse of Little Lost River Valley. In line with it across the Lemhi Range is Pass Creek Pass (Fig. 2 and Pl. V), another deep transverse valley notch of northeast trend. This pass stands out conspicuously from both sides of the Lemhi Range, being stranded 1,000 feet or more above the basins at either end. The head of the stream that carved these now wind-gap valleys has apparently disappeared beneath the surface of the Snake River Plain, but the record of its former northeast trend is indicated by the aligned wind gaps across the Lost River and Lemhi ranges. Its course beyond the Lemhi Range has not been followed.

There are other minor wind gaps that cross the Lost River and Lemhi ranges, and these also afford evidence of a former northeast drainage system. There are also a number of passes that extend across the Beaverhead Mountains, but there was no opportunity in the field to match these with those that cross the Lemhi Range. From southeast to northwest these are the Bannock Pass at 7,672 feet, formerly utilized by the Gilmore and Pittsburgh Railroad; Lemhi Pass at 8,000 feet at the head of Agency Creek; Big Hole Pass at 7,236 feet and just beyond the point where Salmon River turns west; and Gibbonsville Pass at 6,951 feet at the head of North Fork of Salmon River.

Distribution pattern of Tertiary volcanics.—Some of the belts of Tertiary volcanics also indicate an old northeast drainage system. As pointed out by Umpleby[18] and by C. P. Ross,[19] the Tertiary volcanics blanketed a region of considerable relief, in part with as much relief as that which exists today. The volcanics filled the old valleys, and, although the volcanics have been stripped from considerable portions of the region by subsequent erosion, many of the old valley fills still remain and are exposed as belts which may be traced across the area, commonly in a northeast and north direction. Some of the present valleys as well as some of the older wind-gap valleys are directed along these old valley fills. The Salmon River, at least from Challis to and beyond Salmon, is along a belt of volcanics, only remnants of which, in places, remain in upper canyon walls. Another belt extends through Arco Pass, parallel in trend with the present high-level valley. The pre-volcanic valley was considerably broader but little deeper than the re-entrant valley now carved along it. There is another belt of volcanics, through or alongside Wet Creek Pass, which is aligned with patches of volcanics along the south tip of Donkey Hills at the head of Little Lost River Valley and across in the Lemhi Range extending toward the head of Little Lost River. Bannock Pass in the Beaverhead Mountains also is carved in an older volcanic-filled valley. Belts of volcanics extending in a northeasterly direction from Fish Creek and Muldoon to Mackay and vicinity[20] afford additional evidence of an early northeast drainage trend.

[18] Pp. 20 and 34 of ftn. 5.

[19] "Geology and Ore Deposits of the Bayhorse Region Custer County, Idaho," *U.S. Geol. Surv., Bull. 877* (1937), p. 87.

EVIDENCE OF DRAINAGE DIVERSION

Diversion of the early northeast drainage into its present form is indicated by barbed tributaries and elbows of capture and by abrupt changes in direction along

[20] Pl. I of ftn. 5.

fault-made basins in which the streams now occur as consequents.

Salmon River.—Diversion of Salmon River by capture and its virtual reversal of flow is indicated by its sharp elbow turn a few miles below (north of) Salmon and its many barbed tributaries from that point on (Fig. 2). As it makes the turn, it leaves the broad basin of Lemhi Valley, and its direction, which for many miles both in and on the other side of the basin has been north-northeast, changes to northwest and then at North Fork to sharply west. At the same time it enters a profound canyon extending to the Snake River on the opposite side of the state.

Even as the Salmon River begins its elbow turn, its tributaries become barbed, the first, which is Carmen Creek, being almost directly in line with the upper course of the river but with its flow exactly reversed with respect to that of the Salmon. Umpleby[21] apparently noticed this relationship, which prompted him to conclude that the drainage was south until its capture by a vigorous stream from the west. It is possible that Carmen Creek is the inverted course of the northeasterly flowing Salmon River before its capture.

Downstream almost all the tributaries, except the smaller subsequents, are barbed. Chamberlain Creek and Middle Fork of the Salmon River show this relationship exceptionally well (Fig. 2). Panther Creek, however, enters the Salmon at a normal angle, but the alignment of Napias and Moose Creek valleys with upper Panther Creek (Fig. 2) and certain abnormalities of the Napias and Moose Creek valleys suggest that Panther Creek may have been captured by a subsequent tributary of the Salmon and its former course to the northeast left to Napias and Moose creeks. Both Napias and Moose creeks drain old, relatively shallow valleys aligned with upper Panther Creek and are separated from each other by a low valley-floor divide. In flowing southwest, Napias Creek enters a narrow gorge and cascades abruptly into the present valley of Panther Creek. Moose Creek does likewise in the opposite direction, entering Salmon River as a barbed tributary. Traced northeastward, the old valley course of Panther Creek would enter the drainage area of North Fork of Salmon River and possibly would extend through either Big Hole or Gibbonville passes.

Big Lost River.—Some of the headwater tributaries of Big Lost River are controlled by northwest-southeast structural lines, but the main river crosses structural trends and flows northeast until it reaches the broad basin that extends southeast along the base of Lost River Range. It there makes a sharp elbow turn (Pl. I and Fig. 3) and flows down the basin as a consequent stream. That the basin is fault-developed is evident from its relations to Lost River Range. This range has every characteristic of a tilted fault block with the basin on the downthrow side. Unable to continue its course to the northeast across the rising mountain block, the defeated river was diverted southeast along the structural basin. The stranded valley (wind gap) of Double Springs Pass that crosses the Lost River Range remains as evidence of Lost River's former northeast extension and subsequent disruption.

Antelope Creek.—Antelope Creek repeats all the essential features of Big Lost River, except that, as it enters Big Lost River Valley at the base of Lost River Range, it becomes a tributary of Big Lost River. Its course thus changes

[21] Pp. 29–30 of ftn. 2.

abruptly from northeast to southeast as its waters join those of Big Lost River. That its former course extended far to the northeast is indicated by the aligned transverse valleys (wind gaps) across the Lost River and Lemhi ranges. Antelope Creek is thus a defeated stream.

Basin drainage.—Each intermontane basin is deeply floored with gravelly deposits made by streams from the bordering mountains. Some of the streams emerge from valleys cut obliquely across the back slopes of the ranges, apparently retaining previously established northeast drainage courses, but many flow directly down the back slopes. Drainage within each of the basins is by consequent streams. Warm Spring Creek and the Pahsimeroi and Lemhi rivers thus occupy structural basins which drain to Salmon River; Big and Little Lost rivers and Birch Creek occupy structural basins with outlet to the Snake River Plain.

DATING THE DRAINAGE DIVERSION

Since the old erosion surface that forms the summits of the dissected uplands and the back slopes of the tilted block ranges antedates the drainage diversion, it assists in limiting the time at which stream capture and diversion have taken place. As pointed out by Ross,[22] this surface is younger than the deformation of the Challis volcanics (late Oligocene or early Miocene). Assuming that much of Pliocene time might be necessary to truncate the disturbed volcanic strata and older rocks, the planation may not have been completed until well toward the close of the Tertiary period. Drainage until then and later must have continued in a northeast direction.

Capture and reversal of Salmon River probably were accompanied by or closely followed the late Tertiary orogeny when uplift of the region induced vigorous headward erosion of a westward-flowing stream which finally reached back to the main Salmon and diverted it from its northeast course. Block faulting associated with the same orogeny apparently diverted Big Lost River and other northeastward-trending streams into northwest-southeast structural basins. The diversion was completed be ore Wisconsin glaciation in Pleistocene time, for heads of the streams show glacial sculpture and valleys below, and some of the wind gaps contain morainic deposits and outwash trains. Whether the diversions were as late as the early glaciation which antedated the Wisconsin by more than 1,000 feet of canyon cutting[23] is not known. The depth of the wind gaps indicates considerable downcutting before the streams were defeated by faulting. The drainage changes could have been brought about in early Quaternary time. As pointed out by Ross,[24] the record since the formation of the post-Challis surface has been one of intermittent uplift and vigorous erosion, the details of which are not yet fully known. Because of the excellent preservation of the fault scarps, considerable of the movement that delineated the mountain blocks may have taken place in early Quaternary time.

HISTORY OF DRAINAGE DEVELOPMENT

According to Ross,[25] much of the western part of the area (site of the present

[22] Pp. 89–91 of ftn. 19.

[23] C. P. Ross, "Early Pleistocene Glaciation in Idaho," *U.S. Geol. Surv., Prof. Paper 158* (1930), pp. 123–28.

[24] Pp. 91–92 of ftn. 19.

[25] C. P. Ross, "Correlation and Interpretation of Paleozoic Stratigraphy in South-Central Idaho," *Bull. Geol. Soc. Amer.*, Vol. XLV (1934), pp. 937–1000.

Idaho batholith) has been a positive area since about the end of the pre-Cambrian and much of the eastern, a part of the Paleozoic geosyncline that extended alongside. When the region became involved in the late Jurassic orogeny and the intrusion of the Idaho batholith (Cretaceous), it all became land and supplied erosional debris to the neighboring Laramide trough. Thus, a northeastward drainage system apparently had become established even prior to the close of the Mesozoic era. The drainage apparently was maintained through the Laramide orogeny at the close of the Cretacous period, and, as indicated by the distribution of belts of Challis volcanics (late Oligocene or early Miocene) along old northeastward-extending valleys, well into Tertiary time.

The more immediate ancestral drainage probably played an important role in post-Challis planation, and the present northeast drainage pattern may be an inheritance from late Tertiary time. The evidence afforded by headwater streams and by present wind gaps indicates that the drainage was northeast and that the Continental Divide at that time coincided closely with the watershed of the present headwater streams. Except at its very head, where there have been some subsequent adjustments to faulting, Salmon River pursued its present northeast course; but instead of making an elbow turn a short distance below the town of Salmon, it continued on in a northeasterly direction, probably along the present inverted course of Carmen Creek and on toward the present Missouri River. The Salmon was then joined by its East Fork and also by Big Lost River. The latter flowed northeast through the site of Double Springs Pass and entered the Salmon not far below the present mouth of Pahsimeroi River by way of the wind gap across the northwest corner of Lemhi Range. Antelope Creek also flowed in a northeasterly direction through what is now Wet Creek Pass and through the unnamed pass at the head of Little Lost River in the Lemhi Range, but whether it eventually entered the Salmon or continued on as an independent drainage system has not been determined. A stream no longer in existence but which may be referred to as the "Arco River" then flowed northeast through Arco Pass and Pass Creek Pass and probably on across the present Continental Divide. At that time Chamberlain Creek, Middle Fork, and Panther Creek had no local connection with Salmon River but may have joined it some miles beyond the present state line. Thus, the drainage just prior to the crustal disturbance that came at the close of the Tertiary was an integral part of a northeast drainage system that extended on into Montana and presumably to the Arctic or the Gulf of Mexico.

Disruption of this northeast drainage system probably started with the orogeny that came at the close of the Tertiary period. The general uplift apparently favored the vigorous headward erosion of a stream from the west which captured the now barbed tributaries, Chamberlain Creek and Middle Fork, then Panther Creek, and, finally, the main Salmon River. At or about the same time, rising block mountains south of the Salmon River barred the way of other northeast streams, cutting off and diverting one of its main tributaries, Big Lost River, southeast to the Snake River Plain. Antelope Creek also was forced to abandon its northeast course and to become a tributary of Big Lost River. Consequent streams developed within the structural basins, fed by streams from the bordering mountains. Warm Spring

Creek, Pahsimeroi River, and Lemhi River flowed northwest to join the Salmon. Little Lost River and Birch Creek drained southeast along the structural basins to disappear on reaching the Snake River Plain.

More or less closely associated with the mountain uplift was the Snake River downwarp.[26] While the mountains and intervening basins were being arched along an axis parallel to and about 50 miles from the margin of the Snake River Plain, the Snake River downwarp was carrying the ends of the ranges to the level of the plain and then below, where they are concealed beneath a covering of lava that partly filled the great downwarped basin. By the downwarp, the headwaters of Arco River were brought below and became a part of the underfloor of the Snake River Plain and were completely lost from view, with only Arco and Pass Creek passes left as evidence of their former existence.

Since the captures and diversions, the streams have become established in their new courses and are continuing to deepen their valleys in the mountainous areas and to aggrade their floors where they occupy the structural basins.

[26] V. R. D. Kirkham, "Snake River Downwarp," *Jour. Geol.*, Vol. XXXIX (1931), pp. 456–82.

CONCLUSIONS

Piracy and large-scale faulting have had an important role in disrupting a northeast Tertiary drainage system and in fixing the pattern of existing stream systems in the northern Rocky Mountains of east-central Idaho. Piracy has virtually reversed the direction of flow of Salmon River and turned its waters from the Missouri system to the Snake and Columbia rivers and the Pacific Ocean. Block faulting has also diverted some of the tributaries of Salmon River and other streams into structural basins extending southeast to the Snake River Plain. Downwarping involved in the origin of the Snake River Plain has caused the bordering mountains to plunge and disappear beneath the great lava-filled depression and to carry from sight the headwaters of streams that formerly drained to the northeast. As a consequence of these diversions the Continental Divide was shifted about 100 miles east of its location in late Tertiary time.

ACKNOWLEDGMENTS.—The writer wishes to take this opportunity to express publicly his gratitude to Professor O. D. von Engeln, Cornell University, for his helpful suggestions during the preparation of this report.

ACCELERATED CHANNEL EROSION IN THE CIMARRON VALLEY IN SOUTHWESTERN KANSAS[1]

THAD G. McLAUGHLIN
United States Geological Survey, Denver, Colorado

ABSTRACT

The Cimarron River formerly had a narrow channel throughout its course from the Kansas-Colorado state line to the Kansas-Oklahoma state line. Accelerated erosion began in 1914 and has increased the average width of the channel to more than 1,000 feet. Overcultivation appears to have been the principal cause of the erosion.

INTRODUCTION

During investigations of the geology and ground-water resources of Morton, Stevens, Grant, Haskell, and Seward counties, Kansas, for the state and federal geological surveys, the writer has obtained many data indicating extreme widening of the channel of the Cimarron River in southwestern Kansas. The widening was first reported by H. T. U. Smith[2] and was later described briefly by the writer.[3] In most places the accelerated channel erosion began with the disastrous flood of May 1 and 2, 1914, which destroyed many buildings and bridges as well as much of the bottom land. The principal data include testimony of the older residents of the area, photographs taken before the flood of May, 1914, and measurements of the stream channel that were made during a survey of this area in 1874. In addition, data have been obtained from files of newspapers and from histories of southwestern Kansas and of the Santa Fe trail.

The Cimarron River heads in northeastern New Mexico and flows eastward into the panhandle of Oklahoma, turns northeast across the southeastern corner of Baca County, Colorado, and enters Kansas about 5 miles north of the southwest corner of the state. Its course across southwestern Kansas is shown in Figure 1. In southern Meade County the river swings eastward for a short distance and then flows southeastward into Beaver County, Oklahoma, where it flows parallel to the Kansas-Oklahoma line until it re-enters Kansas in south-central Clark County. It then flows eastward into Commanche County, re-enters Oklahoma, and empties into the Arkansas River a short distance west of Sand Springs. The length of the Cimarron is approximately 700 miles, and the length of its course from western Morton County to southern Meade County is 144 miles.

From the Kansas-Colorado line to the first crossing of the Kansas-Oklahoma line the Cimarron River has an average gradient of 8.5 feet to the mile. The gradient is about 9 feet to the mile in

[1] Published with the permission of the director of the United States Geological Survey and of the director of the State Geological Survey of Kansas.

[2] "Geologic Studies in Southwestern Kansas," *Kan. Geol. Surv. Bull. 34* (1940), pp. 174–76; "Notes on Historic Changes in Stream Courses of Western Kansas, with a Plea for Additional Data," *Trans. Kan. Acad. Sci.*, Vol. XLIII (1940), pp. 294–300.

[3] "Geology and Ground-Water Resources of Morton County, Kansas," *Kan. Geol. Surv. Bull. 40* (1942), pp. 14–16; "Geology and Ground-Water Resources of Grant, Haskell, and Stevens Counties, Kansas," *Kan. Geol. Surv. Bull. 61* (1946), p. 36.

Morton and Stevens counties, 7 feet to the mile in Grant, Haskell, and northwestern Seward counties, and 9 feet to the mile in southeastern Seward and southwestern Meade counties. The valley bottom is less than a mile wide in most places but broadens in parts of Seward County and in southwestern Meade County. The width ranges from a little more than 1 mile in southern Grant and southwestern Haskell counties to about 4 miles in southeastern Seward and southern Meade counties. The valley probably is of about the same width in western Morton County, where the south side has a gentle slope that merges with the upland and is masked by deposits of dune sand. Deposits of high-terrace sand and gravel 2.5 miles south of the river indicate that the valley is at least 4 miles wide in that area. The valley is about 100 feet deep between the Colorado-Kansas line and southern Grant County but deepens to more than 250 feet in southeastern Seward and southern Meade counties.

The channel of the Cimarron River is wide and sandy in most parts of southwestern Kansas. The width ranges from about 140 feet at the Satanta bridge in Haskell County (Fig. 1) to more than 3,000 feet in western Morton County. In many places the sides of the channel are inconspicuous, whereas in other places they may be as high as 25 feet. The average height probably is between 5 and 10 feet.

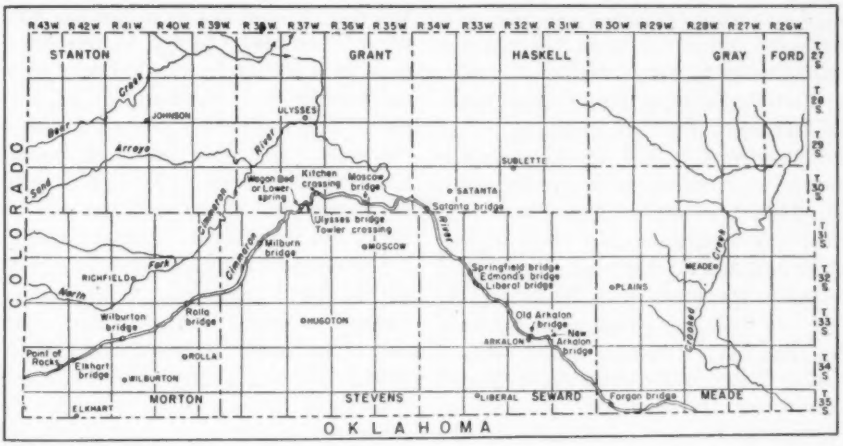


FIG. 1.—Map of southwestern Kansas showing the Cimarron River and its principal tributaries

The principal tributaries of the Cimarron are North Fork Cimarron River and Crooked Creek. North Fork Cimarron, heading in southeastern Colorado, parallels the Cimarron through Morton, Stevens, and Stanton counties into Grant County, where it swings abruptly southeastward and joins the Cimarron near the Grant-Haskell line (Fig. 1). West of this confluence, North Fork Cimarron flows at an altitude lower than that of the Cimarron; hence it probably is the older stream. The course of Crooked Creek, as far as the Oklahoma border, is seen in Figure 1.

GEOLOGY AND GEOLOGIC HISTORY

The part of southwestern Kansas through which the Cimarron River flows is underlain primarily by rocks of Tertiary and Quaternary age. The oldest rocks in the general area are Permian redbeds which crop out along Crooked Creek in Meade County, whereas the oldest rocks in the Cimarron Valley are redbeds which are believed to be Triassic and which crop out in the vicinity of Point of Rocks in western Morton County. Cretaceous deposits underlie the northern part of the area but are not visible in the Cimarron Valley. The oldest Tertiary beds are those that comprise the Laverne formation, cropping out in the Cimarron Valley in southeastern Seward and southwestern Meade counties. They are steeply dipping beds that consist of sandstone and shale in the upper part and thick deposits of sand and gravel in the lower part. They are believed to be Lower Pliocene. Deposits of the Ogallala formation, which are of Middle Pliocene age, crop out in Meade County on the east side of Crooked Creek and in Morton County in the vicinity of Point of Rocks. The Rexroad formation, which is either Upper Pliocene or Lower Pleistocene, crops out in many places in the Cimarron Valley, particularly in Seward and Meade counties, and is overlain by the Pleistocene Meade formation. Between eastern Morton County and the Liberal bridge (Fig. 1) the channel of the Cimarron River is approximately at the level of the contact of the Rexroad and Meade formations. Where the basal sand and gravel of the Meade formation lies in deep channels in the underlying Rexroad, the contact is below the level of the stream bed, and, where the sand and gravel of the Meade are thin, the contact generally is above the level of the stream bed. More recent sediments in this area include dune sand in the Cimarron Valley in Morton and Stevens counties, the alluvium of the Cimarron River, and deposits of colluvium overlying the pediment-like slopes within the valley.

There are at least two and perhaps three terraces in the Cimarron Valley in southwestern Kansas. H. T. U. Smith[4] reported terraces that are 20 and 55 feet above the flood plain in Stevens County and 20, 55, and 80 feet above the flood plain in Seward County. A prominent high-level terrace in Meade County was reported by J. C. Frye and C. W. Hibbard.[5] This terrace has been observed by the writer in many parts of Seward County and in western Morton County. Its level is only about 50 feet below the surface of the upland; hence it probably is the oldest terrace in the valley. Frye and Hibbard report that teeth of *Paraelephas columbi* (Falconer) have been taken from deposits underlying this terrace in Meade County. The remains of fossils and the fact that the terrace deposits are in channels in the Pleistocene Meade formation indicate that this terrace was formed during very late Pleistocene time and that almost all the downcutting in the Cimarron Valley was during later Pleistocene and Recent time. The presence of high-level terrace deposits near the borders of the present valley show that the flood plain during high-terrace time was approximately as wide as the valley now is and was in some places as much as four times as wide as the present flood plain.

[4] Pp. 123 and 126 of ftn. 2 (1940).

[5] "Pliocene and Pleistocene Stratigraphy and Paleontology of the Meade Basin, Southwestern Kansas," *Kan. Geol. Surv. Bull. 38*, Part 13 (1941), p. 420.

STREAM FLOW

Except during floods the flow of the Cimarron River in southwestern Kansas is governed primarily by the relation of the level of the channel to the level of the water table in the upland areas adjacent to the river (Fig. 2). At Point of Rocks the water table in adjacent upland areas is above the level of the stream bed; hence there generally is water in the channel in that vicinity. This point was known as "Middle Spring" on the old Santa Fe trail. From a short distance below Point of Rocks almost to the Rolla bridge (Fig. 1) the water table in areas adjacent to the Cimarron is very slightly below the level of the stream bed. The river in this area is a losing stream, but during periods of excessive precipitation the water table may rise sufficiently to cause a small flow in a few places.

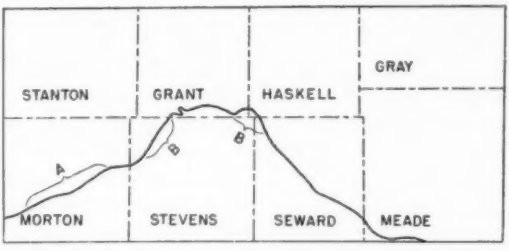


FIG. 2.—Map of southwestern Kansas showing the course of the Cimarron River. *A*, Section in which the water table and river channel are at about the same level. *B*, Sections in which the water table lies below the river channel. The water table in all other sections lies above the level of the stream.

Between the Rolla and the Milburn bridges (Fig. 1) the stream channel is slightly below the level of the water table in adjacent areas, and, hence, there generally is a small flow in that area. This flow disappears rapidly between the Milburn bridge and the Grant-Stevens line, owing to an abrupt reversal of the relationship between the stream channel and the adjacent water table.

For several miles east from the Grant-Stevens line the water table is much higher than the river channel; hence the flow in this section is perennial except perhaps during excessively long droughts. As the water moves eastward, it sinks beneath the channel because the water table becomes lower than the channel. From the Ulysses bridge to northwestern Seward County the channel lies approximately at the level of the contact of the Meade and Rexroad formations. The lower part of the Meade formation consists primarily of unconsolidated sand and gravel, whereas the upper part of the Rexroad formation is of partly consolidated to well-consolidated sand, silt, clay, and caliche. Where the upper part of the zone of saturation is in the lower part of the Meade formation, the water moves freely through the coarse sediments, and the slope of the water table is slight; where the upper part of the zone of saturation is in the upper part of the Rexroad formation, the water moves slowly through the fine-grained sediments, and the water table is steep. In addition, the gradient of the river in southeastern Grant, southwestern Haskell, and northwestern Seward counties is less than in other parts of southwest-

ern Kansas, owing perhaps to the resistance to erosion of the upper part of the Rexroad formation. Because of the increased slope of the water table and the decreased gradient of the river channel, the water table is much lower than the channel in the vicinity of the Satanta bridge (Fig. 1), and, consequently, this section of the river is dry except during floods. In the vicinity of the Springfield bridge, the stream reappears, owing to an increased stream gradient and decreased slope of the water table. The slope of the water table is less in this area because the upper part of the zone of saturation is in the coarse deposits in the lower part of the Meade formation. In southeastern Seward and southern Meade counties the water table is much higher than the river channel, and the flow, therefore, is perennial.

An over-all picture of the stream flow in the Cimarron River is best shown by Figure 3, which is a graph of the stream flow based on measurements at six places in southwestern Kansas. The graph shows a net loss in stream flow between Point of Rocks and the station at Satanta bridge, with the flow generally increasing greatly from the Liberal bridge to the Forgan bridge south of Plains. The loss in stream flow from Point of Rocks to the Ulysses bridge (north of Hugoton) appears to be nearly uniform except during periods of low flow, when the loss is very slight. During periods of low flow the gradient from the surface of the water in the channel to the water table in adjacent areas is small, and little water moves laterally into adjacent areas. When the stream flow is high, the gradient is increased and more water moves into adjacent areas.

The graph also indicates that a flow of nearly 40 cubic feet a second at Point of Rocks is required to cause a flow at

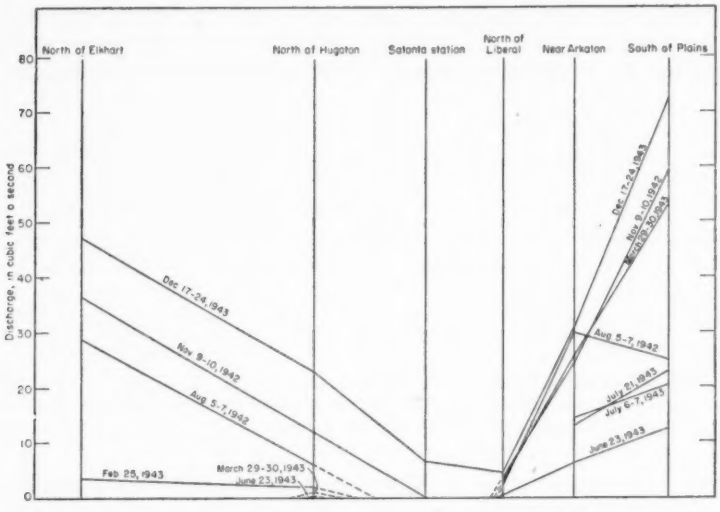


FIG. 3.—Graph of stream flow in the Cimarron River in southwestern Kansas based on measurements made by the United States Geological Survey and the states of Kansas and Oklahoma. (*Kan. Geol. Surv. Bull. 61.*)

the Satanta bridge and that a flow of 45–50 cubic feet a second at Point of Rocks would leave no dry stretches in the channel in southwestern Kansas.

EARLY DESCRIPTIONS

One of the best early accounts of the Cimarron River is that of Josiah Gregg,[6] who traversed the Santa Fe trail several times before 1840. On June 19, 1831, his party reached the Cimarron at a point several miles east of Wagon Bed Spring and found the river dry. They traveled southward for one day and then northwestward until they again encountered the Cimarron (near Lower or Wagon Bed Spring), which "with its delightful green-grass glades and flowing torrent had all the aspects of an 'elysian vale,' compared with what we had seen for some time past." In the same year, when a fur-trader, Jedediah Smith, arrived at Lower Spring, "his eyes were joyfully greeted with the appearance of a small stream meandering through the valley that spread before him."[7] During a later crossing of the Cimarron, Gregg[8] says of it:

> Its water disappears in the sand and reappears again, in so many places, that some travelers have contended that it "ebbs and flows" periodically. This is doubtless owing to the fact, that the little current which may flow above the sand in the nite, or in cloudy weather, is kept dried up, in an unshaded channel, during the hot summer days . . . . In these sandy stretches of the Cimarron, ... travelers procure water by excavating basins in the channel a few feet deep, into which the water is filtrated from the saturated sand.

On June 12, 1847, when Dr. A. Wislizenus[9] reached Lower Spring, he described the Cimarron as "a small green valley, spread out like an oasis in the desert. The water is fresh and running, and the rushes grow on the banks." Ten years later, Eugene Bondel wrote that the Cimarron River above Middle Spring (at Point of Rocks) is "a narrow, running stream, which can be jumped across almost anywhere."[10] Orestes St. John,[11] who made a reconnaissance of the geology of southwestern Kansas in 1885, depicted the river thus:

> [It is a] small brook only a few yards wide, and in places during a portion of the year its waters are lost in the sandy bed. In early summer its low banks are sometimes overflowed. The above-mentioned affluents mostly afford pools the year round, and like the larger stream they are subject to overflow from the heavy local rains that occur during the summer months. In the past these pools were the resort of herds that pastured the adjacent plains, and the Cimarron Valley was, until recently, entirely occupied by stock ranches and thousands of cattle.

Erasmus Haworth[12] stated in 1897 that "the Cimarron seems to have reached base-level and to have begun meandering across its flood plain. Beautiful oxbow curves are frequent, and a sluggish nature is everywhere manifest during times of low water." Five years later (1902), W. D. Johnson[13] wrote:

> There is strong suggestion that by fan construction in the High Plains region it [the Arkansas River] has lifted itself out of a former deep valley and shifted to another course, the

[6] *Commerce of the Prairies* (New York: Henry G. Langley Co., 1844), Vol. I, pp. 74–93.

[7] *Ibid.*, p. 93. [8] *Ibid.*, Vol. II, pp. 197–204.

[9] "Memoir of a Tour to Northern Mexico in 1846 and 1847," *Sen. Misc. Doc. 26* (30th Cong., 1st sess. [1848]), p. 13.

[10] Ralph P. Bieber, *Frontier Life in the Army* (Glendale, Calif.: Arthur H. Clark Co., 1942), p. 179.

[11] "Notes on the Geology of Southwestern Kansas," *Kan. State Board of Agric., 5th Biennial Rept. for Years 1885–86* (1887), p. 133.

[12] "Underground Waters of Southwestern Kansas," *U.S. Geol. Surv. Water-Supply Paper 6* (1897), p. 22.

[13] "The High Plains and Their Utilization," *U.S. Geol. Surv., 22d Ann. Rept.*, Part 4 (1902), pp. 647 and 663–64.

deep valley thus abandoned being that now occupied by the feeble Cimarron. . . . . The valley is . . . . almost unvisited by run-off floods, yet with a perennial stream of constant volume looping intricately upon it. . . . . It does not run full length. Though in some of its live sections it is a strong stream, unvarying in volume, there are other sections in which the bed is permanently dry to depths of 20 or 30 feet. . . . . As elsewhere, under like circumstances, wild hay is cut here for winter feed. . . . . Wherever within the High Plains belt the Cimarron Valley shows a living stream it is always a meandering, looping stream of uniform width, narrow, clear, and deep.

In 1905 W. G. Russell[14] observed that, about 14 miles northwest of Hugoton, the Cimarron "is very sluggish and dirty, with little flowing water, but having in its bed many large pools. The bottoms are sand covered with straggling sage brush and a little grass, with here and there a small body of grass land from which hay is cut." The following year C. N. Gould[15] stated that, in the panhandle of Oklahoma, the Cimarron was no more than 20 feet wide. Later, H. N. Parker[16] wrote:

From the old post office at Metcalf, Okla., to Point of Rocks, Kans., a distance of 25 miles, the channel of the Cimarron is often dry, but at Point of Rocks, Kans., the water comes to the surface . . . . and the channel is usually full for a number of miles. . . . . In Kansas, from Arkalon southward, the Cimarron River usually has water in it throughout the greater part of the year. . . . . William Easton Hutchison states that the Cimarron River is a constantly running stream throughout the entire width of Morton County, where it has a valley on one side or the other of the channel . . . . on which an abundant crop of natural hay is cut. In Stevens County there is running water in Cimarron River at all seasons of the year. . . . . In Grant County the river flows constantly and has a fine fertile valley on each side of the channel that is sometimes covered by floods.

Although the above accounts contain certain minor inconsistencies, the general picture is clear.

RECENT TESTIMONY

Early residents of southwestern Kansas agree that the Cimarron River then had a narrow channel but that its width has increased greatly since the flood of 1914 and continues to increase. In Morton County, for example, E. M. Dean of Richfield stated that the river formerly was a small meandering stream and that abundant crops of natural hay were grown in the flood plain. The width of the channel at Point of Rocks was 66 feet in 1874, whereas in 1916 it was necessary to build a bridge having a length of 644 feet in order to span the channel at that place. The width of the channel in 1939 was approximately 1,400 feet (Pl. I). Several ranch buildings that were situated on the flood plain at Point of Rocks in 1937 had been engulfed by the channel by 1944. Figure 4, *A*, is a view of the Cimarron River channel at Point of Rocks in 1939.

The widening of the channel has been particularly great in some parts of northwestern Stevens County. The history of the channel in sec. 23, T. 32 S., R. 39 W. is typical of those areas where widening has been very great. The width of the channel at the east line of section 23 in 1874 was 36 feet (probably measured in the direction of the line of the survey rather than normal to the direction of stream flow). In 1930, when the Panhandle Eastern Pipeline Company constructed a pipeline across the channel in section 23, the channel had widened to

[14] "Investigations in Kansas," *House Doc. 44* (58th Cong., 2d sess. [1904]), p. 327.

[15] "Geology and Ground-Water Resources of Oklahoma," *U.S. Geol. Surv., Water-Supply Paper 148* (1905), p. 90.

[16] "Quality of the Water Supplies of Kansas," *U.S. Geol. Surv. Water-Supply Paper 273* (1911), pp. 306–7.

PLATE I

Aerial view of Cimarron River in the vicinity of Point of Rocks (lower left corner at *N*) and Elkhart bridge (upper right center). Note narrow patches of alluvium (darker areas) along north side of valley floor that have not yet been destroyed by the widening channel. The channel has almost engulfed the small island of alluvium downstream from Elkhart bridge. Arrow is about a mile long. (Photograph by U.S. Department of Agriculture.)

FIG. 4.—*A*, Cimarron River at Point of Rocks in 1939. (Photograph by H. A. Waite.) *B*, Flood plain of Cimarron River in 1899. Note the narrow, steep-walled channel in the foreground. (Photograph by W. D. Johnson.)

1,200 feet.[17] During the next year the same company drilled a gas well in section 23, and, although the well was located on the flood plain about 300 feet north of the river channel, the widening of the channel was so great that by July, 1942, the well was in the stream bed at a point about 410 feet south of the north bank of the channel (Pl. II). The left wall of the channel, therefore, has been shifted 710 feet in about 11 years.

The widening of the channel has not been so great in Grant and Haskell counties as in Morton and Stevens counties, owing to the narrowness of the flood plain in this section and to the decrease in stream gradient. Mrs. Oscar McClure of Hugoton stated that her father built one of the first bridges across the Cimarron in southern Grant County. The bridge, located southwest of Wagon Bed Spring, was 32 feet long, whereas the channel was about 500 feet wide at that point in 1943. G. E. Gordon of southern Grant County used a 20-foot plank for a foot bridge across the channel of the Cimarron before it increased to its present width of more than 400 feet.

R. R. Wilson[18] of Garden City wrote the following about the Cimarron River:

> I am satisfied that the channel was not over 30 feet wide there [at the Ulysses bridge] in 1886 and was mud bottom. My father built a bridge across the river just a few rods above where the Ulysses bridge is now located in the spring of 1886 and I remember this was a very short bridge. . . . . There was always water in the river from Wagon Bed Spring down stream for about 10 miles, until along about 1897 or 1898 I camped on my homestead about 6 miles below Wagon Bed Spring and that summer was very dry and the water would all disappear about noon and through the afternoon, then every morning there would be a stream of water running. . . . . I am quite sure the channel was always dry at the Satanta bridge . . . . , also the channel was nearly always dry above Wagon Bed Spring for some distance.

Views of the Cimarron River at Towler crossing (site of Ulysses bridge) in 1903 and in 1943 are shown in Plate III.

Judge William Easton Hutchison,[19] of Garden City, writes:

> The bed of the stream there, the Kitchen crossing, was in my best recollection about 10 feet wide, surely not more than 12 feet. There was always water covering the bed of the channel, I would say 6 or more inches deep. . . . . I frequently drove from Ulysses to . . . . Liberal in Seward County, but I kept north of the river and didn't cross it until I reached the Edmonds crossing north of Liberal about 6 or 8 miles. There seemed to be a little more water there than at the Kitchen crossing in Grant County—maybe 10 or 12 inches deep but not much wider than at the Kitchen crossing. I also crossed this stream at Arkalon occasionally. That was very much like the Edmonds crossing.

J. W. Dappert,[20] of Taylorville, Illinois, county surveyor of Grant County in 1887 and 1888, writes as follows:

> I had been up and down the Cimarron River in Commanche County, Kansas and . . . . in the Cherokee strip [northern Oklahoma], where the river was then [1886] from 200 to 600 feet in width with a very sandy bed; but up here [Wagon Bed Spring] we found a little brook with water holes at places 30 or 40 feet wide and 100 to 200 feet long; but between the water holes a narrow channel ran along such as I could and did easily step across, with a small stream of pure sparkling water flowing along in the channel. . . . . Along the Cimarron River grew a sort of slough grass, heavy in foliage, rank and green. . . . . Also flags or cat-tails grew out of the lowest parts of the marsh area of some 20 acres which was a part of the Dick Joyce homestead [NE $\frac{1}{4}$ sec. 33, T. 30 S., R. 37 W.].
>
> Again, on this trip [May 14, 1916], I took special notice of the great changes in the condition and format of the Cimarron River. I had known it when I could easily step across its channel—only 3 or 4 feet wide and 4 or 5 feet

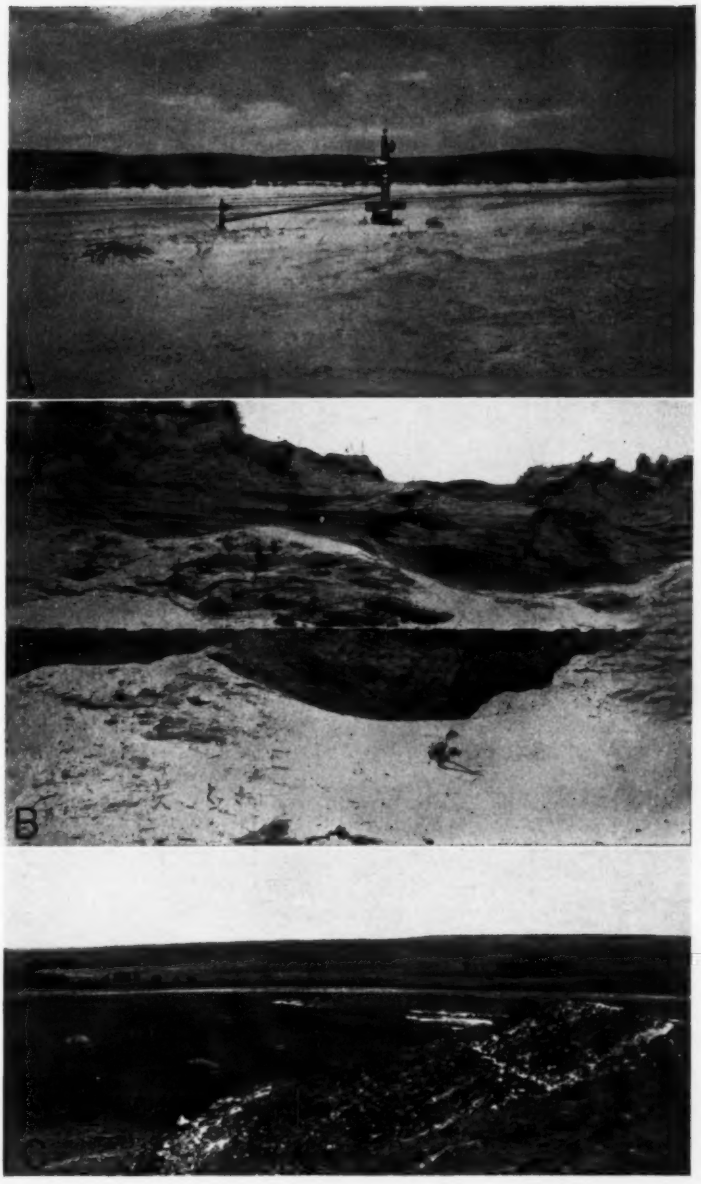
[17] Data from Samuel Neithercoate of the Panhandle Eastern Pipeline Co.

[18] Letter dated April 1, 1943.

[19] Letter dated April 2, 1943.

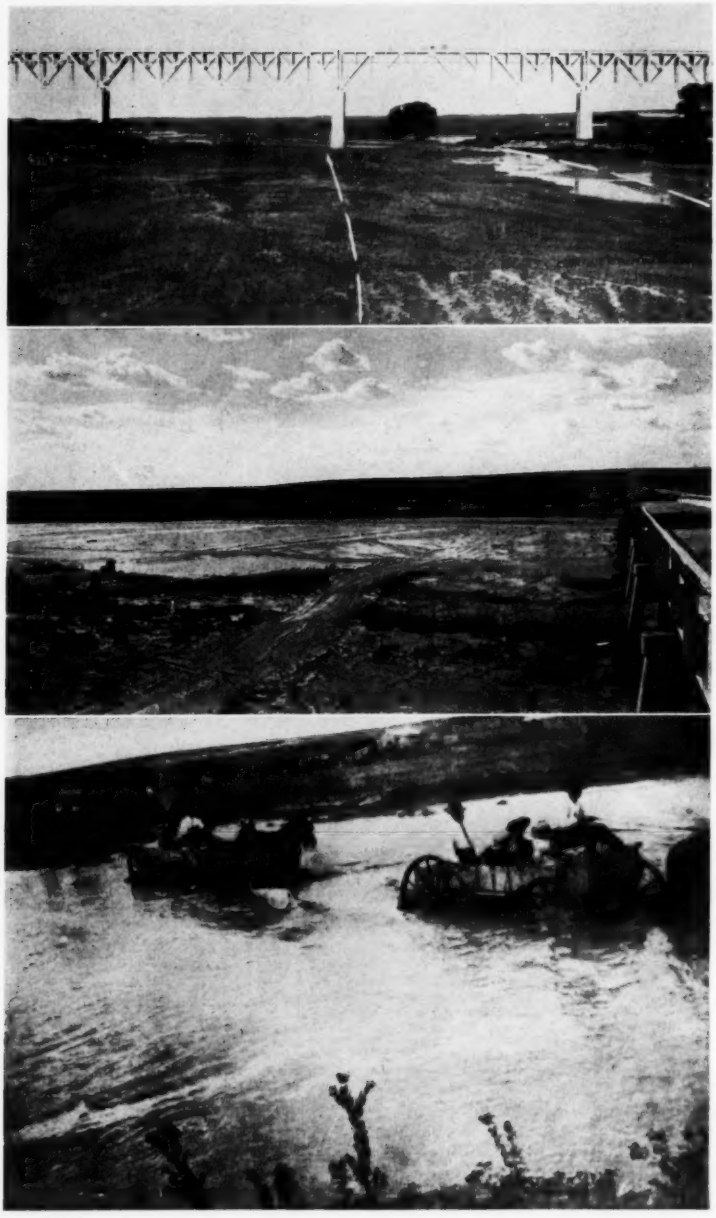
[20] Letter dated January 1, 1944.

PLATE II



A, Gas well in channel of the Cimarron River in sec. 23, T. 32 S., R. 39 W., Stevens County. *B*, North wall of channel of Cimarron River in sec. 23, T. 32 S., R. 39 W., Stevens County. Widening of channel has exposed dune sand overlying alluvium. *C*, Hay field on flood plain of the Cimarron in sec. 25, T. 34 S., R. 31 W., Seward County. This is one of the few places where the hay meadows have not been completely destroyed. Tilted beds of the Laverne formation in the foreground.

PLATE III



A, Cimarron River at New Arkalon bridge in 1943. White lines indicate width of channel in 1939. *B*, Towler crossing in 1943. Ranch buildings are on the brink of the channel wall. *C*, Towler crossing during the flood of 1903. Note gentle slope from ranch buildings to stream. (Photograph by D. D. Crotts.)

deep— . . . . and now, ... by actual measured [by pacing] distance, the width of the sandy bed was 450 feet [near Wagon Bed Spring] . . . . with a few trickles of water visible and flowing feebly along in the sandy bed. . . . . At Wagon Bed Spring the river had receded some hundreds of feet farther to the east, thus shortening the quite abrupt bend. . . . . I made a sketch plat of the stream showing its location and width; and later superimposing my earlier map [1886] upon my later map I noted the difference in the character of the stream. . . . . At first view there was no sand visible at all, while at its later view there was little else visible but sand.

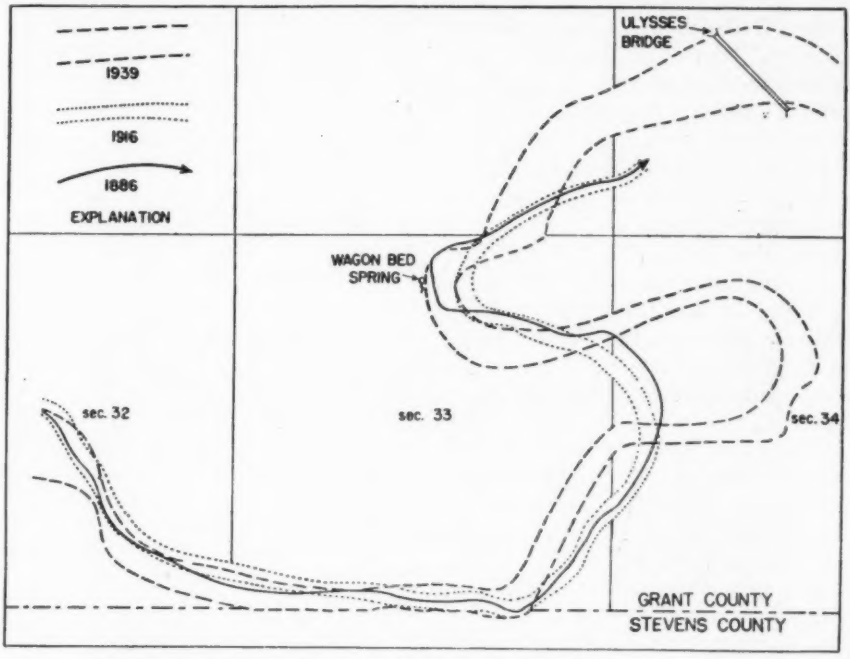


FIG. 5.—Map of a part of T. 30 S., R. 37 W., showing the Cimarron River as sketched by J. W. Dappert in 1886 and 1916 and as taken from aerial photographs made in 1939.

Figure 5 is a map of the vicinity of Wagon Bed Spring showing the position and width of the Cimarron in 1886, 1916, and 1939.

Few data concerning the width of the channel in Haskell County before 1914 are available. The widening of the channel has been less in Haskell County than in any other part of southwestern Kansas, owing perhaps to the position of the water table and the relatively low gradient of the stream. The average width of the channel in that area in 1874 was 32 feet (based on measurements at five places), whereas the width ot the Satanta bridge in 1937 was 64 feet (measured by H. T. U. Smith) and in 1942 was 140 feet.

The destruction of valuable grazing and meadow land by the widening channel of the Cimarron probably has been greatest in Seward and Meade counties, although the widening has been less there than in Morton and Stevens counties. In

Meade and Seward counties the wider flood plain afforded a much greater area of hay meadows prior to the widening. In the vicinity of the Liberal bridge the channel has widened from only a few feet before 1890 (according to Bernard Lemert) to 790 feet in 1939.

T. V. Harvey of Seward County supplied photographs that show the radical changes in the channel between Liberal bridge and Arkalon (Pl. IV). In 1939 the width of the channel at the site of photograph *A* was 1,760 feet and at the site of photographs *B* and *C* was 1,100 feet.

According to Lee Larrabee of Liberal, many acres of bottom land have been destroyed in the vicinity of Arkalon. Photographs taken by him during the flood of May, 1914, show many stacks of hay in the valley bottom (Pl. V) indicating a relatively narrow channel and a broad flat flood plain. Plate V, *C*, shows the broad sandy channel in that vicinity in 1943.

Henry Cochran, foreman of the XI Ranch in Meade County, stated that he had cut hay where there is now only a broad sandy channel. Mr. William Sourbeer reported to John C. Frye of the Kansas Geological Survey that he formerly transported his well-drilling machine across the channel of the Cimarron on two 30-foot planks. The width of the channel at the narrowest point in Meade County in 1939 was about 500 feet.

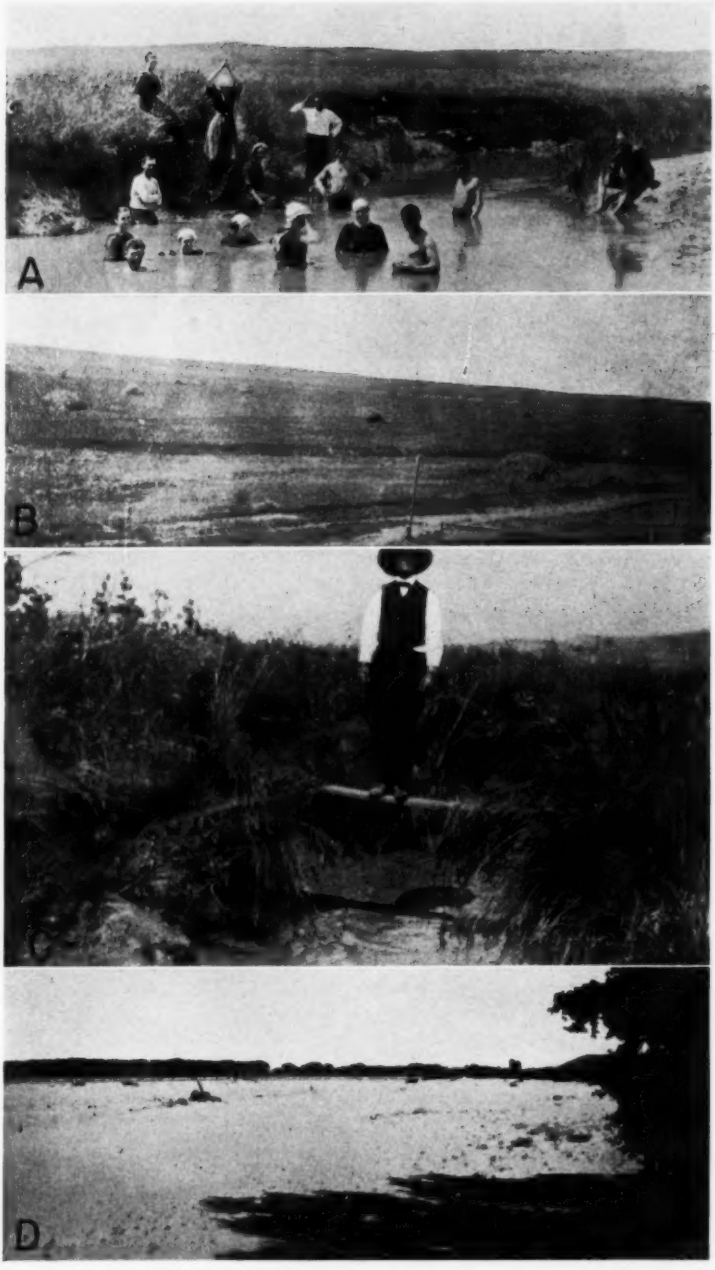
SURVEY OF 1874

The land network of most of southwestern Kansas was surveyed in 1874 during which time the width of the channel of the Cimarron River was measured at almost every section line that crossed the stream. The measurements apparently were made on the section lines rather than normal to the stream flow. This is indicated by the measurement made at the east line of sec. 34, T. 30 S. R. 37 W., Grant County, where travelers of the Santa Fe trail and early settlers in Grant County reported a very narrow channel. The surveyors reported a width of 132 feet, indicating that their measurement was not made normal to the stream which at that point flows almost parallel to the north-south section line. The measurements of 1874 generally were listed to the nearest 5 or 10 links. The greatest width reported by them was 304 feet in southern Meade County, the least was 10 feet in T. 32 S., R. 33 W., Seward County, and the average width was about 51 feet. The channel was widest in the few miles before the river entered Oklahoma. The width in this section (the last two townships through which the river flows in Meade County) averaged 154 feet, whereas above this section it averaged 45 feet. In Table 1 are listed the widths of the channel in 1874 and in 1939. Measurements listed in italics were taken from contact prints of aerial photographs and probably are accurate to within 10 or 15 feet, whereas all other measurements in 1939 were determined from aerial mosaics and probably are accurate only to within 100 feet. The greatest average width in 1939 was about 2,000 feet in Morton County, the smallest average width was about 480 feet in Haskell County, and the average width from the Colorado line to the Oklahoma line was about 1,160 feet (based on 156 measurements).

ACCELERATED EROSION IN TRIBUTARY STREAMS

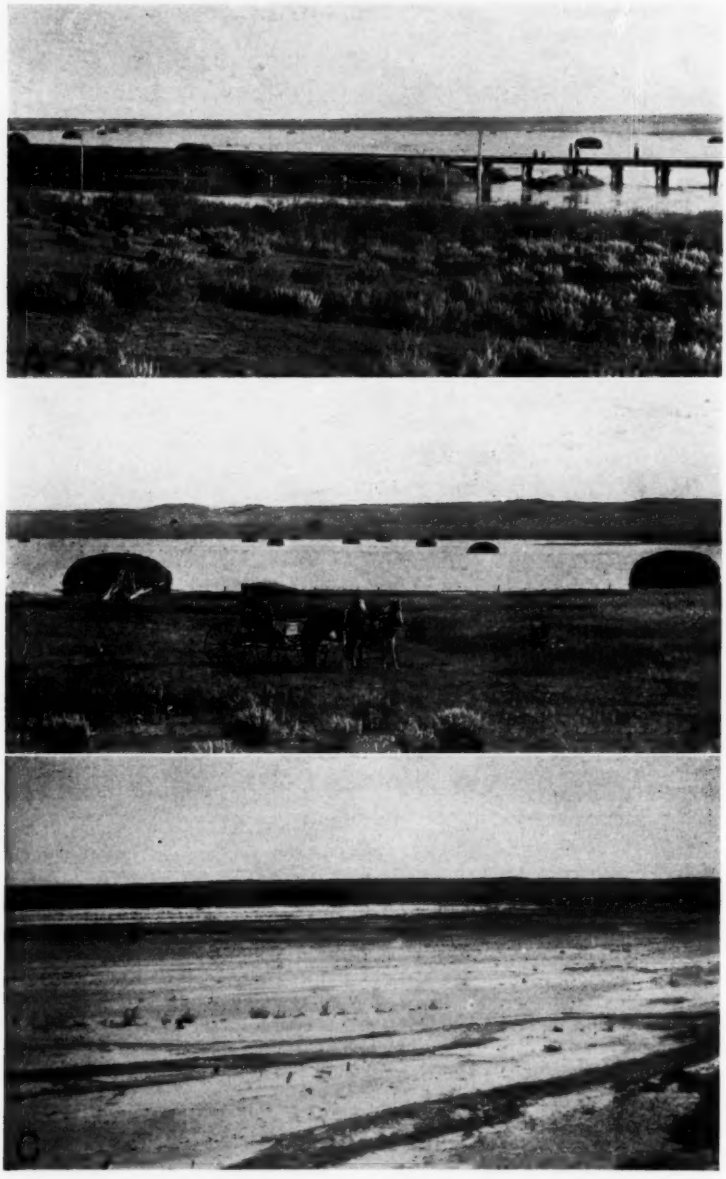
There has been comparatively little channel erosion in the major tributaries of the Cimarron River owing primarily to the lower stream gradients. According to

PLATE IV



A, Channel of Cimarron in sec. 20, T. 33 S., R. 32 W. in 1897. (Photograph by Lee Larrabee.) *B*, View southwestward across Cimarron Valley in NW¼ sec. 18, T. 33 S., R. 32 W. in 1907. Heavy foliage in center marks position of channel. (Photograph by T. V. Harvey.) *C*, Close-up view of channel at a point near the left end of *B*. (Photograph by T. V. Harvey.) *D*, Cimarron channel in 1943. View northward from point about 100 yards downstream from left end of *B*.

PLATE V



A, Cimarron Valley at Old Arkalon bridge during flood of May 2, 1914. (Photograph by Lee Larrabee.)
B, Cimarron Valley about a mile below Arkalon during flood of May 2, 1914. (Photograph by Lee Larrabee.)
C, Cimarron Valley about a mile below Arkalon in 1943.

TABLE 1

WIDTHS OF THE CHANNEL OF THE CIMARRON RIVER IN 1874 AND 1939

Location	Width in 1874 (Feet)	Width in 1939 (Feet)
MORTON COUNTY:		
T. 34 S., R. 43 W.		
West line sec. 19	53	2,900
West line sec. 20	53	3,000
West line sec. 21	46	3,200
West line sec. 22	40	2,800
North line sec. 22	33	2,100
West line sec. 14	79	1,800
East line sec. 14	66	2,300
North line sec. 13	.....	2,400
T. 34 S., R. 42 W.		
West line sec. 7	139	1,300
East line sec. 7	66	*2,150*
North line sec. 8	.....	*2,380*
East line sec. 5	59	*1,400*
East line sec. 4	46	2,600
North line sec. 3	73	1,800
T. 33 S., R. 42 W.		
West line sec. 35	46	3,200
North line sec. 35	46	800
East line sec. 26	46	800
South line sec. 25	66	1,100
East line sec. 36	.....	800
T. 33 S., R. 41 W.		
South line sec. 30	66	1,800
West line sec. 29	66	2,200
West line sec. 28	66	2,600
East line sec. 28	46	1,100
South line sec. 22	66	1,500
West line sec. 23	66	2,000
West line sec. 24	66	2,400
NE corner sec. 24	46	2,500
T. 33 S., R. 40 W.		
East line sec. 18	.....	1,700
East line sec. 18	.....	1,600
East line sec. 18	33	300
South line sec. 8	33	2,100
West line sec. 9	66	2,900
SW corner sec. 3	56	2,400
NE corner sec. 3	43	2,600
T. 32 S., R. 40 W.		
NE corner sec. 35	33	1,100
T. 32 S., R. 39 W.		
West line sec. 30	46	2,600
East line sec. 30	53	1,300
East line sec. 29	53	2,400
Av. in Morton County	56	2,000
STEVENS COUNTY:		
T. 32 S., R. 39 W.		
West line sec. 27	53	2,200
West line sec. 26	46	2,500
North line sec. 26	46	2,100
NE corner sec. 23	36	2,200
North line sec. 13	33	1,800
North line sec. 12	40	700
East line sec. 1	.....	1,900
STEVENS COUNTY—*Continued*		
T. 31 S., R. 38 W.		
South line sec. 31	.....	1,300
SW corner sec. 29	46	1,300
SW corner sec. 21	40	1,700
North line sec. 21	46	1,000
West line sec. 15	40	1,100
NE corner sec. 15	46	800
North line sec. 11	.....	1,000
East line sec. 2	.....	800
North line sec. 1	.....	600
Av. in Stevens County	43	1,400
GRANT COUNTY:		
T. 30 S., R. 37 W.		
West line sec. 32	26	600
West line sec. 33	40	500
West line sec. 34	40	600
South line sec. 28	40	500
West line sec. 27	50	800
North line sec. 34	66	800
West line sec. 35	132	300
South line sec. 26	99	400
South line sec. 23	66	500
West line sec. 24	66	300
North line sec. 25	66	700
NE corner sec. 25	.....	200
T. 30 S., R. 36 W.		
North line sec. 19	.....	500
West line sec. 17	.....	300
North line sec. 20	.....	400
West line sec. 21	66	200
West line sec. 22	66	300
West line sec. 23	66	500
South line sec. 23	.....	300
East line sec. 26	.....	200
East line sec. 25	.....	1,000
T. 30 S., R. 35 W.		
West line sec. 29	33	1,800
West line sec. 28	50	1,100
North line sec. 33	50	1,300
West line sec 34	26	1,300
South line sec. 27	26	1,300
South line sec. 22	33	800
SW corner sec. 23	17	1,100
North line sec. 26	33	700
West line sec. 25	43	500
North line sec. 25	.....	300
East line sec. 24	50	600
Av. in Grant County	52	600
HASKELL COUNTY:		
T. 30 S., R. 34 W.		
North line sec. 30	33	800
East line sec. 30	33	500
South line sec. 29	33	500

TABLE 1—*Continued*

Location	Width in 1874 (Feet)	Width in 1939 (Feet)
HASKELL COUNTY—*Continued:*		
East line sec. 32	26	300
South line sec. 33	33	300
Av. in Haskell County	32	500
SEWARD COUNTY:		
T. 31 S., R. 34 W.		
North line sec. 8	66	300
North line sec. 17	66	300
West line sec. 16	132	500
North line sec. 21	33	600
North line sec. 28	20	1,400
West line sec. 27	66	1,000
North line sec. 34	132	1,000
West line sec. 35	43	1,500
West line sec. 36	33	1,600
South line sec. 36	13	1,000
T. 32 S., R. 33 W.		
West line sec. 6	16	1,100
North line sec. 7	15	1,200
North line sec. 18	16	300
West line sec. 17	16	1,000
North line sec. 20	18	1,000
West line sec. 21	18	600
North line sec. 28	16	1,100
West line sec. 27	16	900
North line sec. 34	13	500
West line sec. 35	10	1,000
West line sec. 36	13	1,300
South line sec. 36	20	500
T. 33 S., R. 32 W.		
West line sec. 6	49	500
North line sec. 7	13	700
North line sec. 18	33	1,800
South line sec. 18	13	440
East line sec. 19	.....	260
East line sec. 20	20	1,410
South line sec. 21	26	370
West line sec. 27	26	790
West line sec. 26	20	1,100
West line sec. 25	33	350
North line sec. 36	33	1,320
East line sec. 36	33	1,180
T. 33 S., R. 31 W.		
South line sec. 31	33	880
SEWARD COUNTY—*Continued:*		
T. 34 S., R. 31 W.		
West line sec. 5	20	1,320
North line sec. 8	.....	910
West line sec. 9	20	270
North line sec. 16	20	1,140
West line sec. 15	26	260
North line sec. 22	26	1,230
SE corner sec. 15	33	970
SE corner sec. 23	30	620
East line sec. 25	33	570
Av. in Seward County	32	900
MEADE COUNTY:		
T. 34 S., R 30 W.		
South line sec. 30	33	1,050
South line sec. 31	79	570
T. 35 S., R. 30 W.		
East line sec. 6	40	1,320
South line sec. 5	43	1,410
East line sec. 8	.....	1,400
South line sec. 9	53	530
East line sec. 16	33	1,250
East line sec. 15	33	800
East line sec. 14	66	1,000
East line sec. 13	132	1,000
T. 35 S., R. 29 W.		
North line sec. 18	.....	1,700
East line sec. 7	.....	1,100
South line sec. 8	.....	1,400
North line sec. 17	132	1,300
East line sec. 8	100	1,100
East line sec. 9	93	1,600
East line sec. 10	145	1,000
East line sec. 11	304	1,300
East line sec. 12	.....	1,300
East line sec. 13	132	800
T. 35 S., R. 28 W.		
South line sec. 18	172	800
Av. in Meade County	99	1,100
Av. through all counties	51	1,160

J. C. Frye,[21] the gradient of Crooked Creek is about 5 feet to the mile upstream from Fowler, about 3 feet to the mile between Fowler and Meade, and about 8 feet to the mile downstream from Meade. Similarly, North Fork Cimarron River has a lower gradient than the Cimarron River; through most of Grant County it is between 5 and 7 feet to the mile. Near its confluence with the Cimarron River, however, its gradient is much steeper, and some channel erosion has occurred there.

Channel erosion has been very extensive in the small tributary streams with-

[21] "Geology and Ground-Water Resources of Meade County, Kansas," *Kan. Geol. Surv. Bull. 45* (1942), p. 14.

in the Cimarron Valley. These streams are most abundant in the wider section of the Cimarron Valley between the Haskell-Seward county line and the Kansas-Oklahoma line. They head in the relatively steep bluffs near the edge of the upland and flow down the broad pediment-like slopes which are covered with deposits of colluvium. These short tributaries were once narrow and shallow, but most of them have been enlarged greatly since 1914. One short tributary (sec. 24, T. 34 S., R. 31 W., Seward County) was formerly crossed by a wagon road and later by an ungraded auto road. In 1943 its channel was approximately 6 feet deep and 50 feet wide.

The channels of these tributary streams appear to have been enlarged headward. One short tributary about 3 miles below the Liberal bridge has an enlarged channel only in the lower half of its course. An auto road that formerly crossed the stream along a section line now crosses at a point about $\frac{1}{4}$ mile upstream, just above the head of the large gully. Along this $\frac{1}{4}$-mile section of the stream are the remnants of several roads each of which began at the old section-line road and crossed the tributary at successively greater distances upstream.

Reports of the early travelers and settlers indicate that the channel of the Cimarron was narrow and shallow. Its walls now are as much as 30 feet high. This, however, is no certain indication that the channel has been lowered. Before accelerated erosion began, the deposits of colluvium covered the pediment-like valley slopes, as well as much of the alluvium, and extended nearly to the Cimarron River. As the channel was widened by cutting laterally into the colluvium, the channel walls could become progressively higher even without downcutting.

Downcutting in the channel of the Cimarron, however, is indicated by the rapid development of tributary gullies and by the upstream migration of the locus of reappearing stream flow below the Satanta bridge. The rapid headward erosion of the short tributary gullies indicates an increased gradient caused by a lowering of the main channel. The reappearance of stream flow below the Satanta bridge was about a mile farther upstream in 1943 than it was prior to 1914.[22] This indicates that the channel is now much lower with respect to the water table than before 1914. Some difference might be caused by a rise in water table, but a relative change in level amounting to about 10 feet cannot be explained solely by a rising water table. Ground-water studies in this area indicate very little change in the water table since 1890 except for seasonal fluctuations. It appears, therefore, that the upstream movement of the point of reappearance of stream flow was caused in part by a lowering of the channel.

CAUSES OF ACCELERATED EROSION

Accelerated channel erosion is common in the arid Southwest, where several detailed studies have been made by C. W. Thornthwaite, C. F. S. Scharpe, and E. F. Dosch[23] and by Kirk Bryan.[24]

[22] Lee Larrabee, personal communication.

[23] "Climate and Accelerated Erosion in the Arid and Semi-arid Southwest, with Special Reference to the Polacca Wash Drainage Basin, Arizona," *U S. Dept. Agric. Tech. Bull. 808* (1942).

[24] Date of Channel Trenching (Arroyo Cutting) in the Arid Southwest," *Science*, Vol. LXII (new ser., 1925), pp. 338–44; "The Papago Country, Arizona: A Geographic, Geologic, and Hydrologic Reconnaissance with a Guide to Desert Watering Places," *U.S. Geol. Surv. Water-Supply Paper 499* (1925); "Channel Erosion of the Rio Salado, Socorro County, New Mexico," *U.S. Geol. Surv. Bull. 790* (1927), pp. 17–19; and "Historic Evidence on Changes in the Channel of Rio Puerco, a Tribu-

These studies show that the greatest incidence of channel erosion was during the period from 1880 to 1900. Enlargement of the channel of the Cimarron River began with the flood of May 1 and 2, 1914. Big Sandy Creek (another tributary of the Arkansas River) has eróded similarly since about the same time, but to a lesser extent.

The cause, or causes, of accelerated valley erosion in the arid and semiarid Southwest are not known, and the theories are conflicting. The rejuvenation of streams could be a possible cause. Local rejuvenation, such as that resulting from stream capture, would not account for the widespread accelerated erosion. A regional uplift would cause rejuvenation of streams flowing down the dip of the beds but would decrease the gradient of streams flowing in the opposite direction.[25]

Proponents of the theory that a change in climate is the cause of accelerated erosion believe that a decrease in precipitation causes the vegetation to be less resistant to run-off, and hence, increases erosion. Others believe that a decrease in precipitation leads to a depletion of vegetation and that the resulting increase of sediment overloads the streams, causing them to aggrade rather than degrade their channels. Thornthwaite, Sharpe, and Dosch point out:

> Decrease of precipitation in an arid or semiarid land reduces the vigor of plant growth. The less hardy specimens die, the survivors are reduced in size or strength, and there is a general shift toward a more xerophytic vegetation. With decreasing precipitation there is also less water to flow off the lands. Three results are possible: (1) If the effect of vegetal depletion is less than that of reduction in precipitation, run-off will be less destructive and the hypothetical change of climate will bring less erosion than formerly; (2) vegetal depletion may balance decrease in precipitation and erosion conditions may remain the same as before; (3) depletion of vegetation may be more effective than the decrease in precipitation and, particularly if individual storm intensity remains the same, run-off and erosion will be accelerated.[26]

Bryan and others believe that a climatic change has been the principal cause of accelereted erosion in the arid Southwest. As evidence, Bryan states:

> On Chaco River . . . . an arroyo has been cut since 1860. Previously, however, an arroyo similar to the modern arroyo was formed in late prehistoric time. This channel was filled with alluvium and the floor of the valley once more became a true flood plain. Obviously the erosion of this prehistoric arroyo could not have been brought about by overgrazing. On Rio Puerco a similar ancient arroyo, completely filled by renewed sedimentation, has been discovered. . . . . Thus on both of these streams cyclic changes from sedimentation to erosion and back to sedimentation took place before the introduction of livestock and the formation of the existing arroyos. It appears inherently most probable that these cyclic changes have a common, and doubtless a climatic, cause. The introduction of livestock and the ensuing overgrazing should be regarded as a mere trigger pull which timed a change about to take place.[27]

Others have observed gullying in places where cattle had never been grazed.

Thornthwaite, Sharpe, and Dosch[28] concluded that the prehistoric channel erosion described by Bryan probably was caused by extreme climatic fluctuations such as could be expected only once in 100 years or perhaps only once in 500 years.

---

tary of the Rio Grande in New Mexico," *Jour. Geol.*, Vol. XXXVI (1928), pp. 265–82.

[25] P. 107 of ftn. 23.

[26] P. 109 of ftn. 23.

[27] P. 281 of ftn. 24 (1928).

[28] P. 46 of ftn. 23.

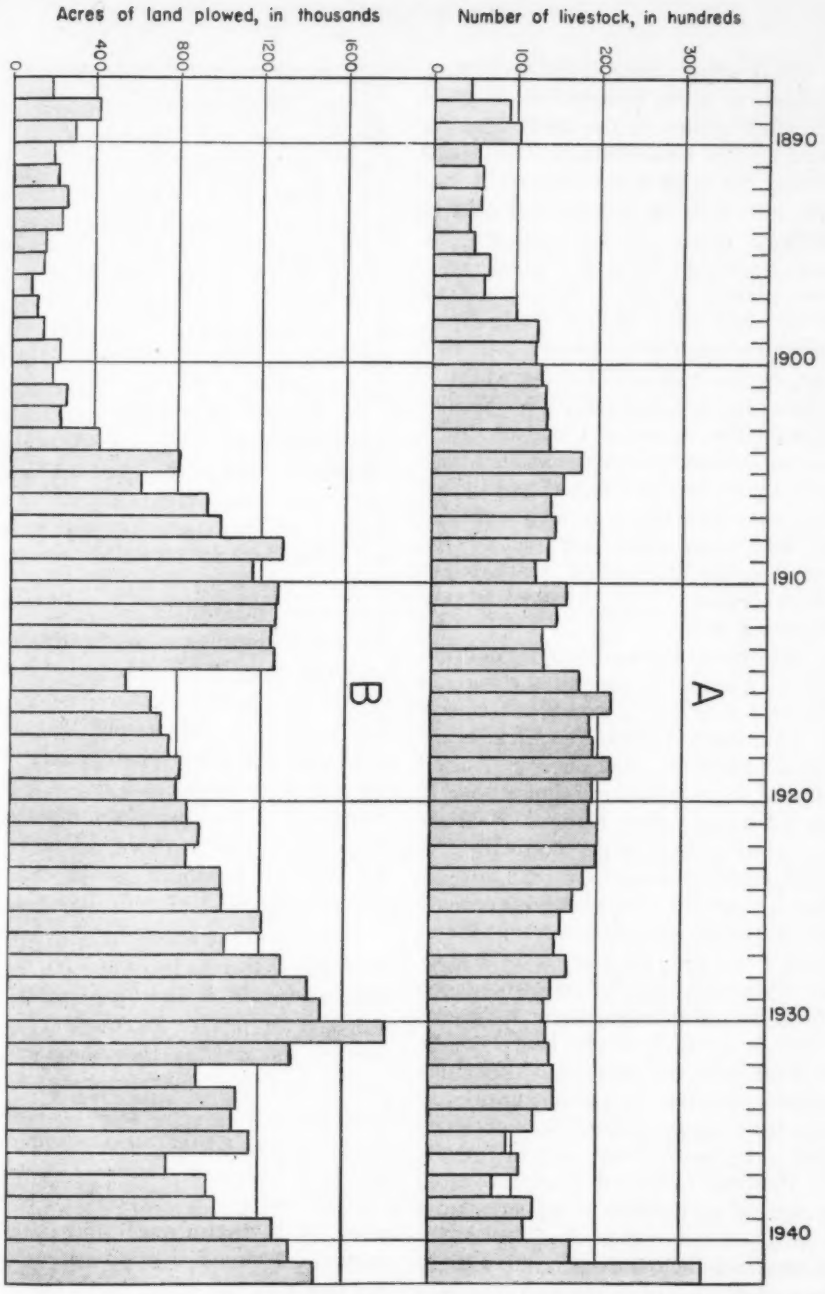


FIG. 6.—*A*, Livestock in southwestern Kansas between 1888 and 1942. *B*, Acres of land plowed in southwestern Kansas between 1888 and 1942.

Overgrazing commonly has been thought to be the principal cause of accelerated erosion in the arid regions of southwestern United States. Accelerated erosion began in many streams in that area soon after the introduction of large herds of cattle which overgrazed the pastures. In the Cimarron Valley, however, grazing appears to have had little or no effect upon erosion. Figure 6, *A*, shows the number of livestock in that part of southwestern Kansas which is drained by the Cimarron River. (Similar data for the rest of the Cimarron drainage basin were not available.) There was no sharp increase in livestock population in or near 1914; hence, it can be concluded that overgrazing was neither the cause nor the "trigger pull" that brought about erosion of the channel of the Cimarron River.

The overcultivation of land is not considered a cause of accelerated erosion in the arid Southwest, which has always been primarily a grazing country. It appears, however, that overcultivation may be the principal cause of erosion in the Cimarron Valley. Figure 6, *B*, shows the acres of land under cultivation in southwestern Kansas from 1888 to 1942. The acreage of plowed land averaged about 250,000 in that area from 1888 to 1903. From 1903 through 1908 the acreage increased rapidly and from 1909 through 1914 averaged about 1,250,000. After 1914 much of the land lay idle without care and without a protective vegetal covering. It was not until 1928 that the acreage of plowed land exceeded that of the period from 1909 to 1914. It is, therefore, believed that the rapid growth of agriculture in this area from 1909 to 1914 and the subsequent abandonment of thousands of farms was the *immediate* cause of accelerated erosion in the Cimarron Valley.

SUMMARY AND CONCLUSIONS

The evidence here presented proves definitely that the Cimarron River formerly had a narrow channel (meandering in some places) in most of its course between the Kansas-Colorado line and the Kansas-Oklahoma line. Below south-central Meade County the channel was much wider as is indicated by the report of J. W. Dappert and by photographs in the library of the Kansas Historical Society which show that the Cimarron had a moderately wide sandy channel in Clark County prior to 1900. Throughout the river's course in southwestern Kansas there were many areas of flat bottom land on which natural hay was grown, but today there are only a few remnants of these hay fields in southeastern Seward and southwestern Meade counties. The reports on the extent of early floods are conflicting. A few reported that the valley was almost never visited by floods, whereas others said that they were common. Early photographs (Pls. III and V) indicate that floods spread over much of the flood plain, but recent floods have been confined primarily to the channel, inasmuch as the channel now occupies most of the valley floor.

The date of the beginning of extensive widening of the channel was May, 1914, when a flood (Pl. V) destroyed much bottom land throughout the course in southwestern Kansas. The widening has continued since that time and was particularly great during floods in 1941 and 1942. By 1943 all the wooden bridges in this area had been destroyed, and the channel had widened beyond one end of many of the newer steel and concrete bridges. A wooden approach to the Arkalon highway bridge had to be built after a flood in 1941, when the channel

widened beyond the end of the bridge. The damage to bridges has been so great that the Chicago, Rock Island and Pacific Railway Company constructed, at a cost of more than a million dollars, a steel span that extends almost entirely across the flood plain and which probably will not be affected by widening of the channel (Pl. III, *A*). The Elkhart, Rolla, Satanta, and Arkalon bridges are the only highway bridges that have not been destroyed by floods.

The data indicate that the channel of Cimarron River has been lowered but the extent of lowering is not known.

Accelerated erosion in the Cimarron Valley is believed to be the result of overcultivation of land and the subsequent abandonment of thousands of farms, leaving large areas with lessened resistance to run-off.

ACKNOWLEDGMENTS.—The manuscript was read critically by S. W. Lohman and H. T. U. Smith of the United States Geological Survey; by John C. Frye and R. C. Moore of the State Geological Survey of Kansas; and by Claude W. Hibbard of the University of Kansas. I wish also to acknowledge the helpful co-operation of county officials and the many pioneer residents of southwestern Kansas who supplied photographs and many other pertinent data.

THE CATOCTIN FORMATION IN CENTRAL VIRGINIA[1]

ROBERT O. BLOOMER AND RICHARD R. BLOOMER
Virginia Geological Survey

ABSTRACT

The Catoctin formation in central Virginia consists of an undetermined thickness of hydrothermally altered andesite classed as propylite. It is underlain by a variable thickness of both metamorphosed and unmetamorphosed conglomerate, sandstone, tuff, and andesite to which the name "Oronoco formation" is assigned. The Catoctin in northern Virginia is overlain by the Lower Cambrian Loudoun formation. In central Virginia, the Unicoi formation, which is at least partly equivalent to the Loudoun formation, crops out along the western flanks of the Blue Ridge. It is not exposed in the Catoctin outcrop belt because of erosion, faulting, or nondeposition. The Unicoi contains conglomerates, sandstones, tuffs, and an andesite flow lithologically like these rocks in the Oronoco and Catoctin formations. Thus, the andesitic effusives and tuffs in the Oronoco, Catoctin, and Unicoi formations are thought to be genetically related. Furthermore, the Catoctin and Oronoco rest upon a surface of major unconformity. This unconformity underlies the Unicoi. The Catoctin and Oronoco are, consequently, thought to be Cambrian in age. In the Western part of the Piedmont province, the Catoctin overlies the Lynchburg formation. The Lynchburg may therefore be partly equivalent to the Oronoco and thus of Cambrian age.

INTRODUCTION

W. B. Rogers[2] was probably the first geologist to recognize (1836) the igneous origin of the rocks later differentiated as the Catoctin formation. He thought that they were intrusive. G. H. Williams[3] first classified them as volcanic. H. R. Geiger and Arthur Keith[4] first applied the name "Catoctin schist" to these rocks in the legend of a sketch map accompanying a paper published in 1891. The formation was first described in detail by Arthur Keith.[5]

The age of the Catoctin has been in dispute since Keith[6] described it as the oldest formation exposed in the pre-Cambrian complex in the Blue Ridge province in northern Virginia. A. S. Furcron[7] essentially agrees with Keith. A. I. Jonas[8] and Jonas and G. W. Stose,[9] however, state that the Catoctin is the youngest pre-Cambrian formation in the Blue Ridge province in northern Virginia.

The writers have made a detailed study of the geology of the Blue Ridge in central Virginia between Buchanan, Botetourt County, and Afton, Nelson County. There are, in the area studied, two major outcrop areas of the Catoctin formation. This paper does not presume to offer conclusions based upon the geology of the entire Catoctin belt. Studies in central Virginia, however, have dis-

[1] Published with the permission of the state geologist of Virginia.

[2] *Geology of the Virginias* (New York: D. Appleton & Co., 1884), pp. 461–62.

[3] "The Volcanic Rocks of South Mountain in Pennsylvania and Maryland," *Amer. Jour. Sci.*, Vol. XLIV (3d ser., 1892), pp. 482–96.

[4] "The Structure of the Blue Ridge near Harper's Ferry," *Bull. Geol. Soc. Amer.*, Vol. II (1891), Pl. IV.

[5] "Geology of the Catoctin Belt," *U.S. Geol. Surv. 14th Ann. Rept.* (1894), Part II, pp. 306–18.

[6] Pp. 302–9, 311–15, 317–18, Pls. XXII and XXIII of ftn. 5.

[7] "Igneous Rocks of the Shenandoah National Park Area," *Jour. Geol.*, Vol. XLII (1934), pp. 405–7; and "Geology and Mineral Resources of the Warrenton Quadrangle, Virginia," *Va. Geol. Surv. Bull. 54* (1939), pp. 31–32.

[8] "Hypersthene Granodiorite in Virginia," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 56–57.

[9] "Age Relation of the Pre-Cambrian Rocks in the Catoctin Mountain–Blue Ridge and Mount Rogers Anticlinoria in Virginia," *Amer. Jour. Sci.*, Vol. CCXXXVII (1939), pp. 579, 582–90.

closed evidence which seems to indicate that the Catoctin formation may be Cambrian in age.

If the stratigraphic relations of the Catoctin can be firmly established, the formation will provide a much-needed key to the interpretation of the geology of the Blue Ridge and Piedmont provinces in Virginia.

DISTRIBUTION

The Catoctin formation is exposed in the Blue Ridge province for about 200 miles from a point south of Carlyle, Pennsylvania, to a point in central Virginia several miles north of James River. Just north of Harpers Ferry the Catoctin belt bifurcates into two roughly parallel belts that trend southwestward into Virginia. The northwestern belt coincides with the summit area of the Blue Ridge. The southeastern belt is topographically expressed by a series of monadnocks, including Catoctin and Southwest mountains, which forms the foothills of the Blue Ridge. The Catoctin formation described in this paper crops out in the southern part of the northwestern belt (Fig. 1).

STRATIGRAPHIC RELATIONS

In the northeastern part of the Buena Vista quadrangle in central Virginia (Fig. 1), the Catoctin is underlain by a formation composed of conglomerate, arkosic sandstone, tuff, and andesite. Southwestward these rocks are metamorphosed and represented by quartzite, phyllite, and slate. This formation, to which the senior author in 1942 assigned the name "Oronoco"[10] from Oronoco Post Office, Amherst County, Virginia, unconformably overlies the pre-Cambrian crystalline basement complex of the Blue Ridge. This complex in central Virginia is composed essentially of a rock classed as hypersthene granodiorite.[11]

The Catoctin formation overlies the Oronoco formation and, where the Oronoco is absent, pre-Cambrian hypersthene granodiorite. Although the Catoctin is not distinctly or consistently stratified in the Buena Vista quadrangle, it is probably structurally parallel to the underlying Oronoco formation. This relationship is indicated by the parallel relationship, in rugged terrain, of contacts between the Catoctin and Oronoco formations and members in the Oronoco formation.

In eastern Augusta County in central Virginia, the lower portion of the exposed section of the Catoctin formation contains two members composed of tuff, conglomerate, and arkosic sandstone that are, respectively, 790 and 740 feet thick. The lithology of these members is essentially the same as the lithology of the Oronoco formation. Southwest of Augusta County no members comparable to those in Augusta County have been found in the Catoctin formation.

The Catoctin belt in Augusta County is thrust-faulted on both sides, so that neither the top nor the base of the formation is certainly exposed. The two members in that area, however, occur near the base of the exposed section. They may be at least partly equivalent to the Oronoco formation exposed to the southwest of the Catoctin section in Augusta County, or they may represent an upper part of the formation cut out by faulting farther to the southwest. The Catoctin may, in places, grade downward into a sedimentary formation

[10] Robert O. Bloomer, "Geology and Mineral Resources of the Buena Vista Quadrangle, Virginia," *Va. Geol. Surv.* (MS in files).

[11] Geologic map of Virginia, *Va. Geol. Surv.*, 1928.

(Oronoco) through the alternation of altered effusives and sedimentary members or upward into an overlying sedimentary formation (Unicoi) in the same manner. Exposures revealing these relations, however, do not occur in central Virginia.

The Catoctin in northern Virginia is overlain by the Lower Cambrian Loudoun formation,[12] which is at least in part equivalent to the Unicoi formation of central and southern Virginia. In central Virginia, however, the Unicoi rests upon the pre-Cambrian basement complex and not on the Catoctin. The Oro-

[12] *Ibid.*

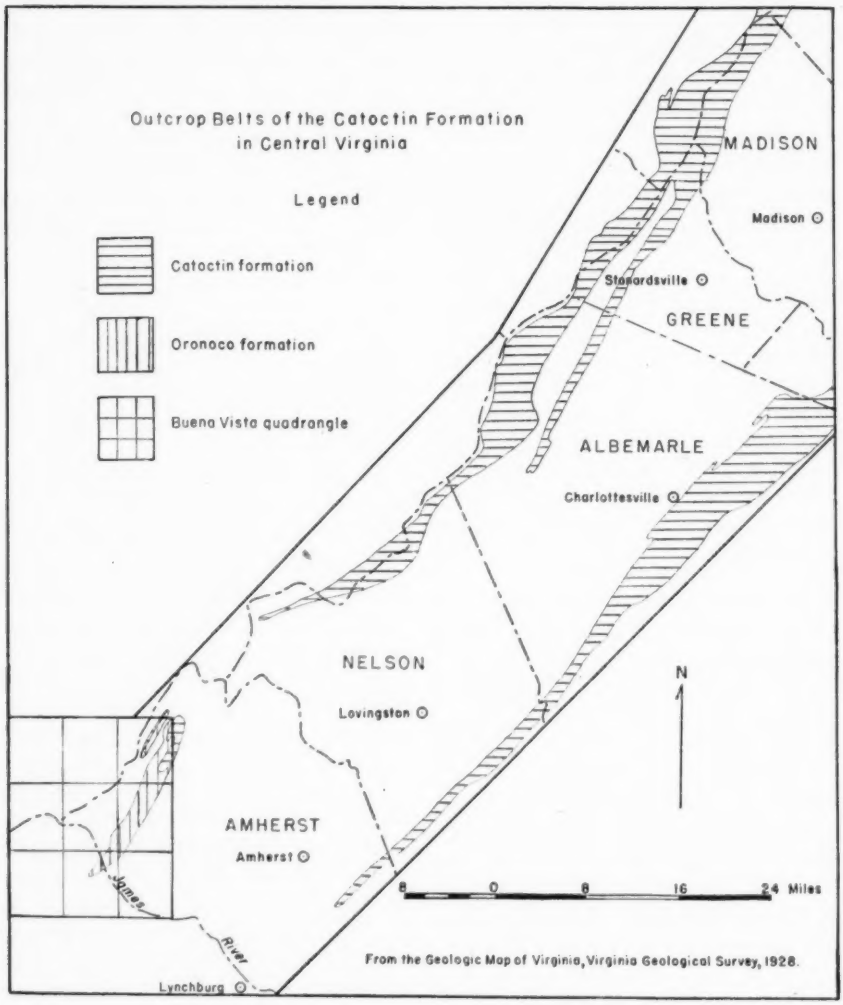
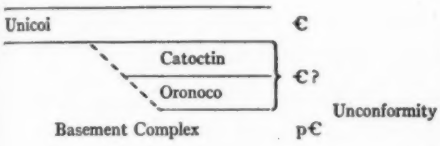


FIG. 1

noco formation, which underlies the Catoctin in central Virginia, rests on the pre-Cambrian crystalline complex except where it is overthrust upon the Unicoi. The Oronoco thins northwestward until the Catoctin rests directly on the pre-Cambrian crystalline complex. Presumably the unconformity beneath the Unicoi decreases in magnitude toward the southeast as the Catoctin and Oronoco come into the section. There is thus an overlapping relationship between the Oronoco, Catoctin, and Unicoi formations (Fig. 3).

TABLE 1

STRATIGRAPHIC RELATIONS OF THE CATOCTIN FORMATION IN CENTRAL VIRGINIA



STRUCTURE

At the southwestern end of the northwestern outcrop belt, in Amherst County, the Catoctin formation is exposed in an anticlinal block bordered on the southeast and on the northwest by overthrust faults. The fault along the northwestern edge of the Catoctin block passes beneath the fault along the southeastern edge just south of James River where the outcrop belt of the Oronoco formation is terminated. Granodiorite[13] forms the core of the anticlinal block, while the Catoctin and Oronoco formations are exposed on the limbs of the fold which plunges northward. A few miles north of James River the Catoctin outcrop belt ends, where it marks the nose of the faulted anticline.

Northeast of the Amherst County area, in Augusta County, the Catoctin is exposed in a block between the Blue Ridge fault and another overthrust fault. In this block, the Catoctin dips toward the southeast and thus represents a part of one limb of a large fold.

The Catoctin belt in central Virginia is bordered on the southeast by the Blue Ridge fault. This fault forms the northwestern margin of the Catoctin Mountain–Blue Ridge anticlinorium[14] which has consequently been thrust upon the Catoctin belt.

PETROGRAPHY

Megascopic character.—The Catoctin formation, as defined in this paper, is predominantly a fine-grained, holocrystalline, green, yellow-green, blue-green, gray, and red-brown altered effusive rock.[15] The structure is massive, except locally where it is slaty or mylonitic. The massive variety is complexly jointed and, in places, includes veins, veinlets, and podlike and irregular masses of milky or clear quartz, pink feldspar, and epidote. This vein material has undergone severe deformation in foliated varieties of the Catoctin formation. In these varieties, chlorite generally incases and obscures introduced aggregates of quartz, feldspar, and epidote.

The texture of the Catoctin propylite is generally uniformly fine grained, but porphyritic varieties with phenocrysts of altered euhedral feldspar occur locally. The texture observed in hand specimens is almost wholly secondary. It consists of an equigranular overgrowth of secondary minerals that obscures the primary tex-

[13] *Ibid.*

[14] Pp. 576–77 of ftn. 9.

[15] There are two members in the unit tentatively identified as the Catoctin formation in Augusta County. These are composed of sedimentary rocks and will be described in a detailed report being prepared for the Virginia Geological Survey.

ture. Incipient weathering clearly reveals the porphyritic character of some of the rock in which, on unweathered surfaces, the phenocrysts are obscured by secondary minerals.

Stratification is not clearly discernible in the Catoctin formation except where thin amygdaloidal and scoriaceous layers are present. It is presumed that the surfaces of these layers represent the surfaces of flows. The amygdaloidal and scoriaceous layers, unfortunately, are discontinuous and only sparsely developed. Hence they do not aid much in determining the attitude of the formation.

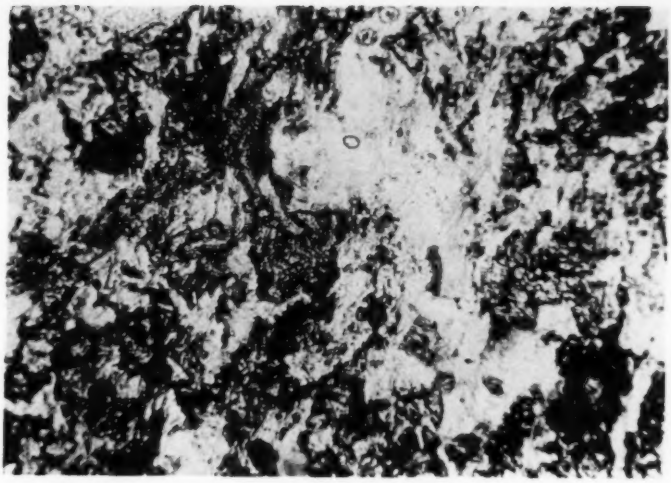
The amygdules in the Catoctin propylite range from about 3 to 15 mm. long. They are generally elliptical and contain pink feldspar, milky quartz, and epidote in all proportions. Other minerals in the amygdules include hematite, chlorite, chalcopyrite, malachite, and zeolite. In the slaty variety the amygdules are represented on S-planes by elliptical or ribbon-like smears dominantly composed of sericite and comminuted quartz.

With the exception of the minerals in the amygdules and rare phenocrysts, the only important rock-forming minerals recognized in hand specimens are epidote and chlorite. Minute grains of a black mineral, probably ilmenite, are discernible in some samples. Epidote appears to be very abundant. It forms an overgrowth that masks most of the primary grains and completely replaces others. The rock consequently appears monomineralic. In some specimens, chlorite can be detected in small micaceous grains subordinate to epidote.

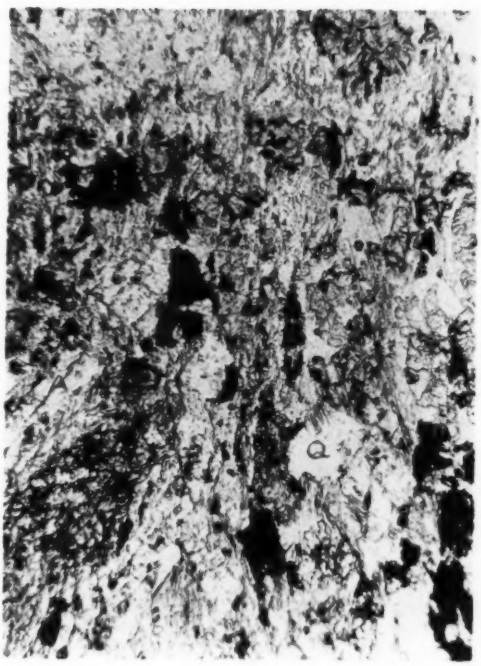
Microscopic character.—Thin sections of the Catoctin effusive reveal a primary texture composed of laths of feldspar that average about 0.3 mm. in length. Phenocrysts of subhedral feldspar, rare in most specimens, range from 1 to 20 mm. in length. Small feldspar laths produce an ophitic fabric with the interstices occupied by fine-grained aggregates of secondary minerals. Secondary minerals also form discrete grains up to several millimeters in diameter, veinlets, saussuritic overgrowths, and embayments in primary grains.

Andesine, quartz, and orthoclase comprise the primary or magmatic minerals discernible in most thin sections. Andesine occurs in small laths. It forms about 50 per cent of the primary mineral composition of the rock. All the feldspar has been altered. Orthoclase forms tiny anhedral grains in the interstices and scanty subhedral phenocrysts. Many orthoclase grains have doubtlessly been completely replaced. Quartz forms minute grains associated with the secondary aggregate and rare larger grains in the interstices. Some of the quartz is secondary or introduced, but the largest grains are probably primary. Some aggregates of chlorite and epidote contain shredlike remnants of biotite. Scanty, minute grains of hornblende and pyroxene (augite?) have also been detected. It is believed, however, that the varietal femic mineral was biotite associated with accessory pyroxene. The hornblende is probably secondary. The recognition of biotite as the varietal mineral is strengthened by the occurrence of biotite in a comparatively thin, much less altered, effusive in the Oronoco formation 330 feet below the base of the Catoctin formation on the west side of Long Mountain in Amherst County. This effusive contains the same essential minerals as the Catoctin propylite. Furthermore, biotite comprises the femic constituent of an effusive in the base of the Unicoi formation (Cambrian) which is otherwise petrographically similar to the

PLATE I



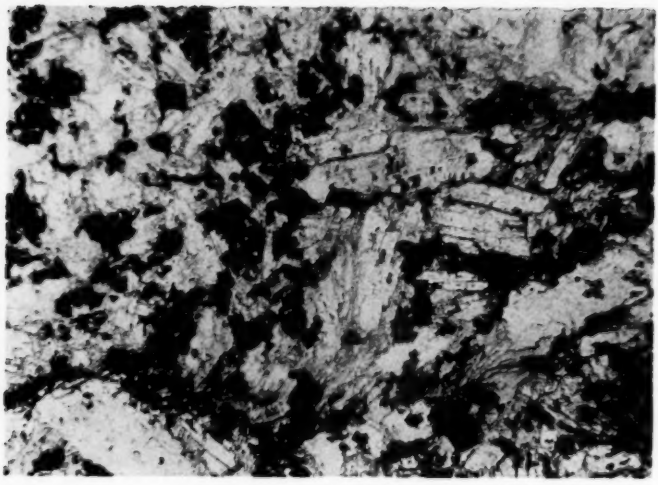
A



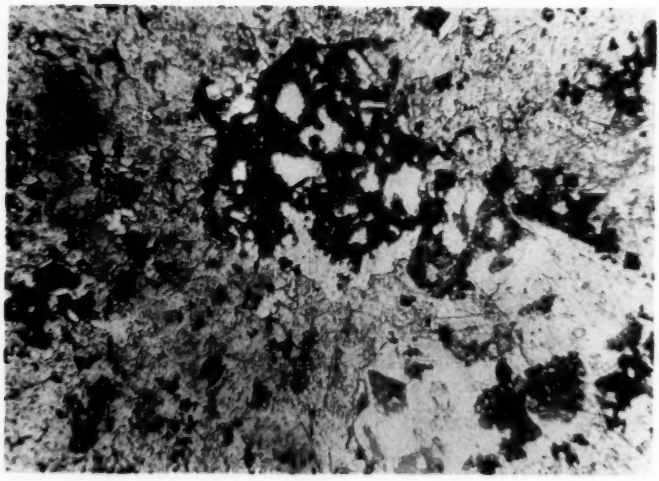
B

A, Catoctin propylite showing an altered phenocryst of orthoclase (*O*), ghost-like laths of andesine, and a heavy overgrowth of secondary minerals. × *ca.* 100. *B*, Catoctin propylite showing vague outlines of andesine laths (*A*), quartz (*Q*) partially replaced by tremolite, and overgrowths of epidote, leucoxene, tremolite, and other secondary minerals. × *ca.* 100.

PLATE II



A



B

A, Propylitized andesite from the Oronoco formation. × *ca.* 100. *B*, Propylitized andesite from near the base of the Unicoi formation. × *ca.* 100.

Catoctin propylite. The effusives in the Oronoco and Unicoi formations are thought to be genetically related to the Catoctin propylite.[16] Some of the most abundant secondary minerals, such as sericite, epidote, chlorite, and iron oxide in the propylite were partly formed from biotite, of which tiny remnants remain. Biotite is also a prominent constituent of some of the dike rock in the granitic basement complex. These dikes are probably the conduits through which the Oronoco, Catoctin, and Unicoi lavas were supplied.

Some of the amygdules in the Catoctin propylite are filled with glass that contains spherulites and is devitrified on peripheries and along cracks. It is also partially replaced by secondary minerals, especially epidote and sericite. No glass has been identified with certainty in the groundmass of the rock. It may be present in minute grains, or it may have been destroyed by devitrification and replacement. The alteration products of the glass observed in amygdules are the same as those of andesine, orthoclase, and biotite and are, consequently, not distinctive.

With the possible exception of andesine, the most abundant minerals in the Catoctin propylite are secondary. These minerals occur as peripheral replacements and overgrowths on the primary minerals which generally appear as shadowy or ghostlike relicts in the groundmass of the rock. The primary features, except for amygdules and rare phenocrysts, are detectable only in thin sections. Even in some thin sections the primary characteristics are obliterated by secondary structures, textures, and minerals. The secondary minerals, in the approximate order of abundance, are: epidote, chlorite, sericite, tremolite, magnetite, ilmenite, hematite, calcite, quartz, leucoxene, and pyrite. The order of abundance and the proportion of the secondary minerals, however, are not invariable. The yellow-green propylite contains a preponderance of epidote. Chlorite is about as abundant as epidote in the blue-green variety. Tremolite is most abundant in the gray variety. A distinctive red-brown variety, so far observed only at the base of the formation in Amherst County, contains a preponderance of iron oxide, including, especially, specularite (Fig. 2).

CLASSIFICATION

Keith[17] recognized andesite, rhyolite, and basalt in the volcanic rocks that comprise the Catoctin formation in northern Virginia. Furcron[18] describes metabasalt, agglomerate, tuff, and sandstone among the rocks in the "Catoctin series" in the Warrenton area, Virginia. In central Virginia the Catoctin is an altered effusive thought to have been andesite or granobasalt. It is, however, difficult to recognize some of the primary characteristics of the rock because of the severe secondary alteration. The texture is clearly fine grained; andesine is the essential mineral; and biotite, orthoclase, and quartz seem to comprise the varietal minerals. Hence, the rock was probably andesite before it was altered.

The Catoctin formation is generally classified as "greenstone," a metabasalt.[19] This classification is somewhat confusing because it is also used to refer to soapstone, metagabbro, and other

[16] T. L. Watson and J. H. Cline, "Extrusive Basalt of Cambrian Age in the Blue Ridge of Virginia," *Amer. Jour. Sci.*, Vol. XXXIX (4th ser., 1915), p. 669.

[17] Pp. 302–9 of ftn. 5.

[18] Pp. 19–28 in *Bull. 54* of ftn. 7.

[19] See ftn. 11.

rocks. The Catoctin is essentially propylite. Perhaps this term should be used in preference to the more general term "greenstone."

Keith[20] also classified the Catoctin as a schist. In places the Catoctin is thinly foliated, but the writers have not seen specimens that could be classified as true schist. The foliated phase is too fine grained to permit the megascopic identification of essential micaceous minerals. Consequently, the foliated Catoctin propylite is perhaps more correctly classified as slate and, locally, as phyllite.

FIG. 2.—Ferruginous foliated phase of the Catoctin propylite showing laths of sericitized andesine. The opaque mineral is chiefly specularite. × *ca.* 100.

ORIGIN

The Catoctin formation in central Virginia is believed to be an altered plateau andesite that emanated from many fissures now represented by propylite dikes in the pre-Cambrian granitic complex of the Blue Ridge province. The rock in these dikes is like the rock in the Catoctin formation. None of these dikes has been found intruding the propylite flows. Lead-uranium ratios from the Catoctin formation and the more or less associated propylite, metagabbro, and metapyroxenite dikes, as well as from the effusives in the Oronoco and Unicoi formations, might disclose approximately equivalent ages and thus indicate a genetic relationship. If such a relationship were established, the widespread dikes and other bodies of propylite in the Piedmont and Blue Ridge provinces would assist greatly in the establishment of stratigraphic and age relations.

There is little doubt of the hydrothermal character of the solutions that altered the Catoctin andesite to propylite. The secondary minerals epidote, sericite, chlorite, calcite, and others and the absence of critical high-temperature minerals suggest that the altering solutions were epithermal. The source of these solutions, however, is problemati-

[20] P. 308 of ftn. 5.

cal. In the Blue Ridge and western part of the Piedmont province, where the Catoctin propylite is extensively exposed, there are no known large, intrusive masses of post-Catoctin age that could have supplied the solutions that altered the formation. Furthermore, with a few local exceptions, the rocks upon which the Catoctin formation rests, including arkosic sandstones, siltstones, tuffs, and granitic rocks, are not similarly altered. No veins or dikes that could have supplied the solutions that so extensively propylitized the Catoctin formation have been found in these underlying rocks. Hence, it is probable that the altering solutions were deuteric[21] solutions in the Catoctin andesite. These solutions probably were retained within the flows by the development of chill-crusts on the lower and upper surfaces. Locally, where cracks were formed in the basal crust, solutions may have escaped and produced hydrothermally altered zones in the underlying rocks. In places, lithified zones in the flows may have been broken or fractured by the movement of the molten effusive. These fracture zones within the flows may have become channels which, when filled with deuteric solutions, formed the veins and masses of quartz, orthoclase, and epidote observed in the Catoctin formation. Some of these veinlike masses containing approximately equal amounts of coarse-grained pink feldspar and quartz resemble pegmatite and granite. The field relations, mineral assemblage, and paragenesis, however, indicate the hydrothermal origin of these masses. Epidote is associated in all proportions with the quartz and feldspar and is contemporaneous with these minerals. There are, furthermore, no critical magmatic minerals. The same mineral assemblage is found in the amygdules.

[21] P. 315 of ftn. 5; and p. 24 in *Bull. 54* of ftn. 7.

THICKNESS

The thickness of the Catoctin formation has not been accurately measured. Such a measurement is difficult, if not impossible, because the exposed sections are faulted at the top or bottom, or at both the top and bottom, and no flow boundaries can be identified with certainty within the propylite. In Amherst County where the base of the section is exposed but the top faulted, the Catoctin is represented by about 2,200 feet of propylite. Northward, in Augusta County, the Catoctin is exposed between faults so that the section is incomplete. There are, however, in the Augusta County section, about 5,000 feet of propylite and sedimentary rocks. The sedimentary rocks occur in the lower part of the section where they form two mappable units (members?), respectively 740 feet and 790 feet thick. In northern Virginia, Keith[22] assigned a thickness of only 1,000 feet to the Catoctin formation. Furcron[23] described an agglomerate

TABLE 2

GENERALIZED SECTION OF THE CATOCTIN FORMATION FROM EXPOSURES IN AMHERST COUNTY, VIRGINIA

	Thickness (In Feet)
Unicoi formation	
Catoctin formation (2,200 feet, estimated)	
4. Propylite, uniformly fine grained, green, yellow-green, gray; recurrently amygdaloidal, massive; locally foliated	1,550
3. Propylite, porphyritic, fine grained, green, massive	100
2. Propylite, fine grained, green, massive	500
1. Slate, red, light-green; flattened, sericitized amygdules	50
Oronoco formation and granodiorite	

[22] Pl. XXIII of ftn. 5.

[23] P. 22 in *Bull. 54* of ftn. 7.

member at least 1,500–2,000 feet thick at the base of the Catoctin in Fauquier County in northern Virginia. Furcron, however, gives no estimate of the total thickness of the formation.

AGE

Keith[24] expressed the opinion that the Catoctin formation is the oldest rock in the Blue Ridge of northern Virginia. He stated that the Catoctin is intruded and metamorphosed by pre-Cambrian granitic rock (granite and granodiorite) but offered no concrete proof of the post-Catoctin intrusion of pre-Cambrian plutons.

Furcron[25] agrees with Keith with regard to the pregranitic complex age of the Catoctin. He reached this conclusion because of an intrusion of "pegmatite" in the Catoctin and because of exposures of "numerous alternations of granite and greenstone [Catoctin] layers." The "pegmatite" mentioned by Furcron is assumed by him to be genetically related to the pre-Cambrian granitic complex in the Blue Ridge. However, as previously stated, these pegmatite-like bodies are probably epithermal in origin and genetically related to the andesite effusives from which the Catoctin formation was formed. Some of these bodies containing quartz and orthoclase in variable proportions closely resemble some pegmatites or granites. Epidote is present in nearly all these bodies and seems to be contemporaneous with the quartz and feldspar. Furthermore, such bodies have not been found in rocks older or younger than the Catoctin formation in central Virginia. These observations and the absence of mica and critical high-temperature minerals suggest that the so-called "injections" are epithermal veins and masses formed from residual solutions in the Catoctin effusives. With regard to the "alternations of granite and greenstone layers," Furcron assumed that the "granite" is the younger rock but states that it has not been found cross-cutting the propylite ("greenstone"). He recognized, however, a body of marble, slate, schist, and gneiss beneath the Catoctin to which he assigned the name "Fauquier formation."[26] Hence, Furcron does not recognize the Catoctin formation as the oldest rock in the pre-Cambrian complex of the Blue Ridge region of northern Virginia as did Keith. The relation of the Fauquier formation and the pre-Cambrian granitic complex is not described.

Jonas[27] and Jonas and Stose[28] are of the opinion that the Catoctin is the youngest pre-Cambrian formation in the Blue Ridge. This conclusion is based upon the absence of pre-Cambrian intrusions in the Catoctin formation, the presence of clastic sediments in the Catoctin derived from the granitic complex, and the extrusion of the Catoctin upon the eroded surface of the granitic basement complex.

The stratigraphic relations and lithology of the Oronoco, Catoctin, and Unicoi formations in central Virginia are indicative of the Lower Cambrian age of these formations.

The Catoctin overlies the Oronoco formation and (where the Oronoco is missing) the pre-Cambrian granitic complex (granodiorite). The Oronoco formation contains conglomerate, tuff, arkose sandstone, and an andesite effusive. Pebbles, cobbles, and boulders of granodiorite

[24] Pp. 302, 309, 311–15, 317–18, Pls. XXII and XXIII, of ftn. 5.

[25] Pp. 405–10 in *Jour. Geol.*, and pp. 31–32 in *Bull. 54* of ftn. 7.

[26] Pp. 37–41 in *Bull. 54* of ftn. 7.

[27] Pp. 56–57 of ftn. 8.

[28] Pp. 579–90, 592 of ftn. 9.

from the pre-Cambrian granitic complex form the conglomerates in the Oronoco formation which concordantly underlies the Catoctin formation. The Oronoco is thus younger than the pre-Cambrian complex and older than the Catoctin formation.

In central Virginia, no rocks resting upon the Catoctin formation are exposed. Farther north, however, the Catoctin is unconformably overlain by the Loudoun formation.[29] In central Virginia the Unicoi formation, at least in part equivalent to the Loudoun formation, is not underlain by the Catoctin but rests unconformably on the granitic complex. The basal portion of the Unicoi contains tuff, an andesite effusive (locally propylite), and a conglomerate composed of cobbles and pebbles from the granitic complex. Propylite fragments are extremely rare or absent in this conglomerate. If the absence of the Catoctin formation beneath the Unicoi formation is due to erosion, the clastic rocks in the Unicoi should, it seems, contain fragments of propylite. It is consequently assumed that, in central Virginia, the Oronoco and Catoctin formations were developed mainly in a basin east of the area of the present Unicoi outcrop belt along the west flanks of the Blue Ridge and never spread westward over the rocks that now underlie the Unicoi. Such an Oronoco-Catoctin basin to the east would explain the absence of the Catoctin beneath the Unicoi formation but would not account for the sparsity of propylite fragments in the Unicoi conglomerates. The erosion of the Catoctin and westward transportation of the sediment would have supplied the Unicoi with propylite fragments. Hence, at least during the first stages of deposition, the Unicoi sediments may have come from a crystalline area to the west of the present outcrop belt of the Unicoi in central Virginia. This area may now be buried beneath Paleozoic sediments in the Appalachian trough west of the Blue Ridge. On the other hand, overthrust faults have moved the present outcrop belt of the Catoctin westward so that it is exposed much nearer the present outcrop belt of the Unicoi formation than it was during the time of Unicoi deposition There may have been an intervening area, not covered by Catoctin propylite, sufficiently wide to permit much dilution and attrition of the Catoctin-derived sediment. This may account for the sparsity of propylite fragments in the Unicoi conglomerates exposed along the west flanks of the Blue Ridge in central Virginia.

The clastic sediments of both the Unicoi and Oronoco formations were derived from the pre-Cambrian granitic complex and are very similar in appearance. Furthermore, the tuffs and andesites in the two formations are very similar in color, texture, mineral composition, and alteration. In fact, without the aid of field relations it is difficult, if not impossible, to distinguish the rocks in the Oronoco formation from those in the lower portion of the Unicoi formation.

The Unicoi is conformably overlain by the Hampton formation, which in turn is conformably overlain by the fossil-bearing Lower Cambrian Erwin formation. This relation points to the Cambrian age of the Unicoi formation.

The Catoctin seems to have been formed from an andesite of the same composition and texture as the andesites in the Oronoco and Unicoi formations. This and the presence of lithologically similar red and gray tuffs in the Oronoco and Unicoi formations are indicative of a period of volcanism that began during

[29] See ftn. 11 and pp. 42–43 in *Bull. 54* of ftn. 7.

Diagrammatic Representation of the Stratigraphic Relations of the Unicoi, Catoctin and Oronoco Formations

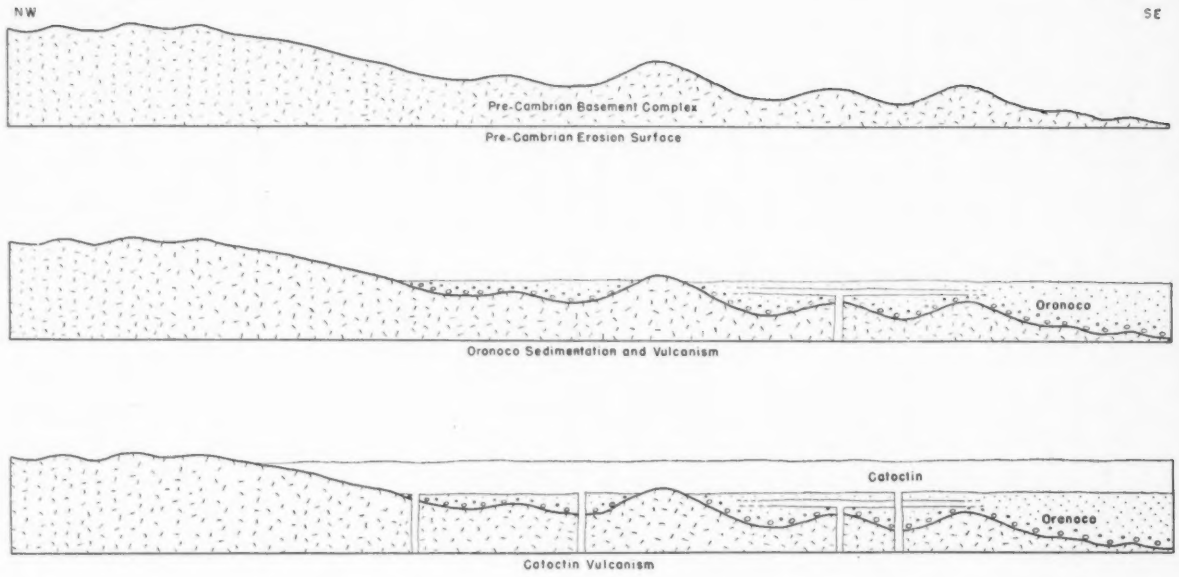
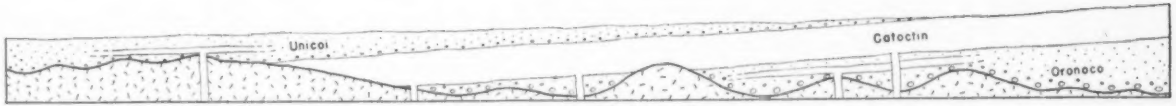
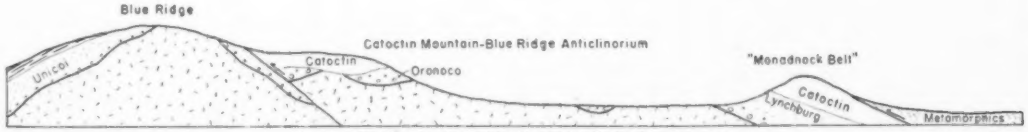


FIG. 3

Westward Tilting and Unicoi Sedimentation Initiated



Unicoi Vulcanism and Sedimentation



Idealized Structure Section of the Blue Ridge North of James River
Showing Present Topography

Oronoco time, continued throughout Catoctin time, and waned during the early part of Unicoi time.[30] Perhaps this volcanism accompanied the initial crustal deformation that resulted in the formation of the Appalachian geosyncline and is consequently one of the first events of Paleozoic time.[31]

The similarity of the lithology and the possible genetic relationship of the volcanics in the Oronoco, Catoctin, and Unicoi formations suggests that these formations were formed under similar conditions. Each of these formations rests upon the eroded surface of the pre-Cambrian granitic complex in the Blue Ridge. In places the Catoctin formation is separated from this significant unconformity by the Oronoco formation, but the Unicoi in central Virginia rests directly upon it. The absence of the Catoctin and Oronoco formations beneath the Unicoi formation in central Virginia probably is due to the relief of the erosion surface on the pre-Cambrian granitic complex (Fig. 3).

The relations and the lithologic similarity of the Oronoco, Catoctin, and Unicoi formations lead to the conclusion that these formations are a continuous stratigraphic series. If the Unicoi formation is Cambrian, then the Catoctin and Oronoco formations are also Cambrian.

The Oronoco formation has not been identified beyond the confines of the Buena Vista quadrangle in central Virginia. Its relation to other sedimentary and metasedimentary formations beneath the Catoctin is problematical. Jonas and Stose,[32] however, have stated that the Lynchburg gneiss (metasediment) is part of the younger pre-Cambrian series that contains the Catoctin formation in the upper part. This suggests that the Oronoco formation, which is overlain by the Catoctin, may be at least partly equivalent to the Lynchburg formation. The Lynchburg formation may then be Cambrian in age. This formation and the Catoctin formation are extensively exposed along the western margin of the Piedmont province in Virginia. The establishment of the age of these formations relative to known Paleozoic formations would greatly aid in interpreting the geology of the Piedmont and Blue Ridge provinces.

ACKNOWLEDGMENTS.—These geologic studies of the Blue Ridge in central Virginia were made for the Virginia Geological Survey. The writers appreciate the suggestions and aid of members of the survey staff. They are especially grateful to Arthur Bevan, W. R. Brown, R. S. Edmundson, and Mr. and Mrs. G. W. Stose for critically reading the manuscript.

[30] See ftn. 16.

[31] G. F. Laughlin, "Comments on the Origin and Major Structural Control of Igneous Rocks and Related Mineral Deposits," *Econ. Geol.*, Vol. XXXVI (1941), p. 674.

[32] Pp. 589–90 of ftn. 9.

A FAULT-PLANE CAVERN

E. L. KRINITZSKY
Southwestern Louisiana Institute

ABSTRACT

The Goodwins Ferry Cave is a linear horizontal cavern having two accessible levels. The linear nature of the cave is due to development on the plane of the Saltville fault, while an earlier control of the drainage by the New River caused the cave to develop horizontally. Enlargement has taken place by solution and caving, and the cave's water is thought to come from overlying sandstones.

LOCATION

The Goodwins Ferry Cave[1] is located in the folded Appalachian Mountains near Goodwins Ferry in Giles County, Virginia, about 25 miles northwest of Blacksburg. The entrance to the cave is at an elevation of 2,015 feet on the end of Spruce Run Mountain, sometimes called Cave Hill, and about 235 feet above the road through the New River gorge.

GEOLOGY OF THE CAVE AREA

I. STRATIGRAPHY

The rocks exposed at Spruce Run Mountain, in the vicinity of Goodwins Ferry, include beds of Ordovician and Silurian age against which Cambrian formations have been thrust along the well-known Saltville fault. The columnar section of these strata is shown in Table 1.

LOWER CAMBRIAN

Honaker dolomite.—The Honaker formation is a thick-bedded, fine-grained, grayish, nonfossiliferous dolomite. No limestone was found at any of the outcrops. The dolomite in some places has been moderately fissured, but the fissures are tightly sealed by veins of dolomite crystals. Near the fault the beds are crenulated with minor folds.

Nolichucky shale.—The Nolichucky formation, of fossiliferous shale and dolomitic limestone, is well exposed about $\frac{1}{4}$ mile south of Goodwins Ferry along a cut of the Virginian Railway. Here a gradational contact with the Honaker dolomite is exposed. The Nolichucky shale dips south at about 55°.

TABLE 1

System	Formations	Thickness (Feet)
Silurian	Clinton ss	80
	Cacapon ss	120
	Clinch ss	55
Ordovician	Juniata sh	95
	Martinsburg sh-ls	800
	Eggleston ls	400
	Moccasin ls	60
	"Lowville" ls	150
	Benbolt ls	100
	(SALTVILLE FAULT)	
Lower Cambrian	Nolichucky dol-sh	55
	Honaker dol	800

ORDOVICIAN

Benbolt limestone.—The Benbolt formation is composed of a dark-blue, fossiliferous limestone alternately thin- and moderately thick-bedded. The most outstanding characteristic is that the beds are made up of small nodules of limestone 1–2 inches in diameter. These nod-

[1] The cavern is generally called the Goodwins Ferry Cave because of the near-by settlement, but it is also known as the New River Cave and, infrequently, as Spruce Run Cave.

ules are believed to result from the jointing and weathering of thin layers of limestone having thin partings of calcareous shale.[2] The Benbolt limestone is not very well exposed on the Goodwins Ferry side of the river, but on the opposite slope, along the cut of the Norfolk and Western Railroad, there is an excellent outcrop showing beds intensely folded near the fault contact with the Honaker dolomite. On the Cave Hill, near the contact with the Honaker dolomite, the limestone has been deformed only by gentle, wavy folding. The contact with the Martinsburg shale represents faulting, but the fault is obscured by soil mantle, and its true nature is not readily determinable. About ½ mile north of the cave the formation has excellent outcrops on both sides of the New River gorge and shows folding with the overlying "Lowville."

Identification of the Benbolt limestone was made by Dr. Byron N. Cooper on fossil evidence collected by the writer. Heretofore, the presence of the Benbolt formation on the Saltville fault contact has not been recognized, and it is the author's belief that the formation here has always been erroneously included with the overlying dolomite. As a result, the plane of the Saltville fault has been mistakenly placed along a minor fault associated with the major thrust.

"Lowville" limestone.—The "Lowville" formation, also best exposed along the railway cuts, is a dark bluish-gray, fine-grained, compact, nonfossiliferous limestone. The formation is generally thin- to medium-bedded, with the thicknesses averaging between 1 foot and 2½ feet, although a few beds run around 6–8 inches. There are several chert horizons near the top of the formation.

[2] Charles Butts, "Geology of the Appalachian Valley in Virginia," *Va. Geol. Surv. Bull. 52* (1940), Part I, p. 171.

Moccasin limestone.—The Moccasin formation is dark-pink, though in places gray-buff, argillaceous limestone, in which deformation has developed pronounced fracture cleavage that gives its outcrops a slatelike appearance. The formation is compact and medium-bedded, the beds averaging from 1 foot to 1½ feet in thickness. The formation is well defined, continuous, and somewhat resistant to weathering, forming a slight ridge in places.

Eggleston limestone.—Moderately folded with the Moccasin limestone is the overlying Eggleston formation, which is a thin- to medium-bedded, fine-grained, argillaceous, buff limestone characterized by a basal chert conglomerate and cuneiform jointing perpendicular to the bedding. The formation contains several beds of bentonite, and the peculiar jointing may be related to this material. The basal conglomerate suggests a disconformity.

Martinsburg shale.—The Martinsburg formation is predominantly a yellowish to brownish shale with numerous limestone beds 1–4 inches in thickness. Very thick beds of clay shale are numerous and quite fossiliferous. Where the shale is thin-bedded, it weathers to an easily recognizable flaky talus. The only good outcrop noted is along the road cut, ½ mile north of Goodwins Ferry, and here only a small part of the formation is exposed. Elsewhere the shale is covered with a soil mantle that obscures it almost completely.

Juniata shale.—The Juniata formation is composed mainly of bright-red shale and beds of brown to red sandstone. The shale is compact, unfossiliferous, and covered with a soil mantle so that there are very few outcrops.

SILURIAN

Clinch sandstone.—The Clinch is a sandstone composed almost entirely of quartz sand and is generally light gray to white. It is a medium-textured, compact, vitreous rock that is practically a quartzite. Its resistant nature determines numerous outcrops mostly as ledges on steep slopes. The sandstone is thick-bedded, with many ripple-marked bedding planes. At the base of the formation, for a thickness of some 30 feet, are conglomeratic layers containing quartz pebbles up to $\frac{3}{4}$ inch in length.

Clinton sandstone.—The Clinton formation is also a compact, thick-bedded, and well-jointed sandstone. The lower half of the formation has highly ferruginous sandstone beds of distinctive character, termed the Cacapon division. The Cacapon sandstone is moderately thin- to thick-bedded and has a striking purplish to brownish-red coloration owing to its iron content. The Clinton formation above the Cacapon division is predominantly a white to gray, medium-textured sandstone having plentiful outcrops, especially along steep slopes.

II. STRUCTURE

The major structural feature of the Goodwins Ferry area is the Saltville or Bland fault extending 200 miles from the north end of Sinking Creek Valley (Craig County, Va.) southwest through Bland and Saltville, and well into Tennessee. The maximum stratigraphic displacement occurs in Washington County, where the Lower Honaker of Cambrian age is brought up against the Mississippian Pennington formation—a stratigraphic distance of 14,000 feet.[3]

On Spruce Run Mountain the Lower Cambrian dolomite is in contact with the Silurian sandstones with a stratigraphic displacement of 4,800 feet. The Honaker dolomite has been thrust from the southeast up against the section which includes the Ordovician and Silurian beds already described. These beds have been bent into a syncline, and the southern limb of the fold has been pinched off by a minor fault which has brought a sliver of the Benbolt limestone into contact with the other Ordovician beds as high as the base of the Clinch sandstone (Fig. 1). The syncline appears to be overturned toward the north and is most intensely folded near the fault.

The Saltville fault, on the slope above the cave entrance, has a dip of 30°, which, below the entrance, rapidly increases to around 50° and is 54° at the foot of the hill. The strike of the fault is, in general, N. 65° E. The minor fault probably has the same strike, but its dip is more uniform, being around 52°.

Along the plane of the Saltville fault the beds of the Honaker dolomite dip southward at the same angle as the fault and then curve upward again to form a syncline whose axis is in the Goodwins Ferry hollow and parallel to the strike of the major fault. South of the syncline the dolomite is in the form of an anticline of about the same magnitude as the syncline and is overlain on the south limb by the Nolichucky shale.

The Goodwins Ferry Cave

I. CAVE DESCRIPTION

The Goodwins Ferry Cave has two predominant levels (Fig. 2). The older, upper level is, in many places, 75 feet in maximum width, while the lower level, where the stream flows, is comparatively small as befits its recent development.

The entrance to the cave is on the fault plane, and both levels of the cave extend

[3] *Ibid.*, p. 457.

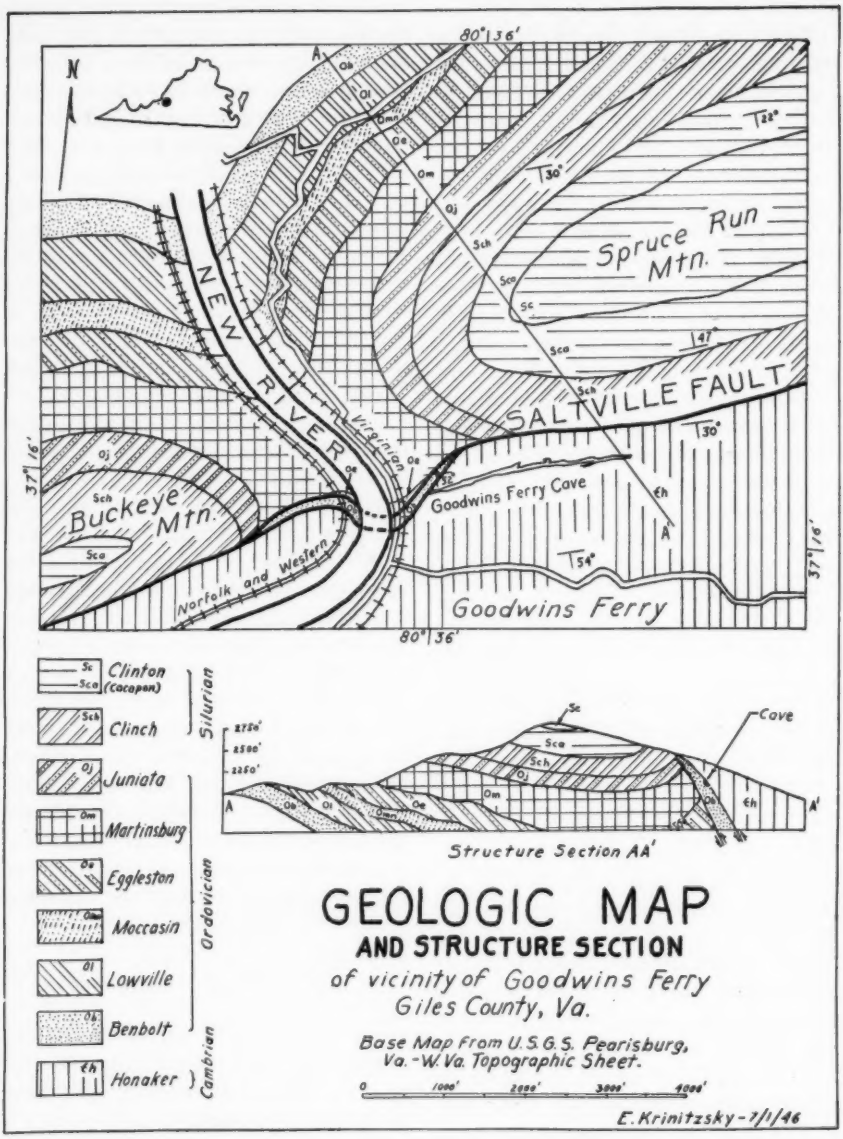

80°|36'
N
A
Ob
Oj
Om
Oe
T22°
T30°
Spruce Run Mtn.
Sch
Sco
Sc
T41°
SALTVILLE FAULT
NEW RIVER
Virginian
Buckeye Mtn.
Goodwins Ferry Cave
Eh
A'
T54°
Norfolk and Western
Goodwins Ferry
37°|16'
Sc Clinton (cacapon)
Sco
Sch Clinch
Silurian
Oj Juniata
Om Martinsburg
Oe Eggleston
Ordovician
Moccasin
Ol Lowville
Ob Benbolt
Eh Honaker
Cambrian
2750'
2500'
2250'
Cove
Structure Section AA'
GEOLOGIC MAP
AND STRUCTURE SECTION
of vicinity of Goodwins Ferry
Giles County, Va.
Base Map from U.S.G.S. Pearisburg, Va.-W. Va. Topographic Sheet.
0 1000' 2000' 3000' 4000'
E. Krinitzsky - 7/1/46

FIG. 1

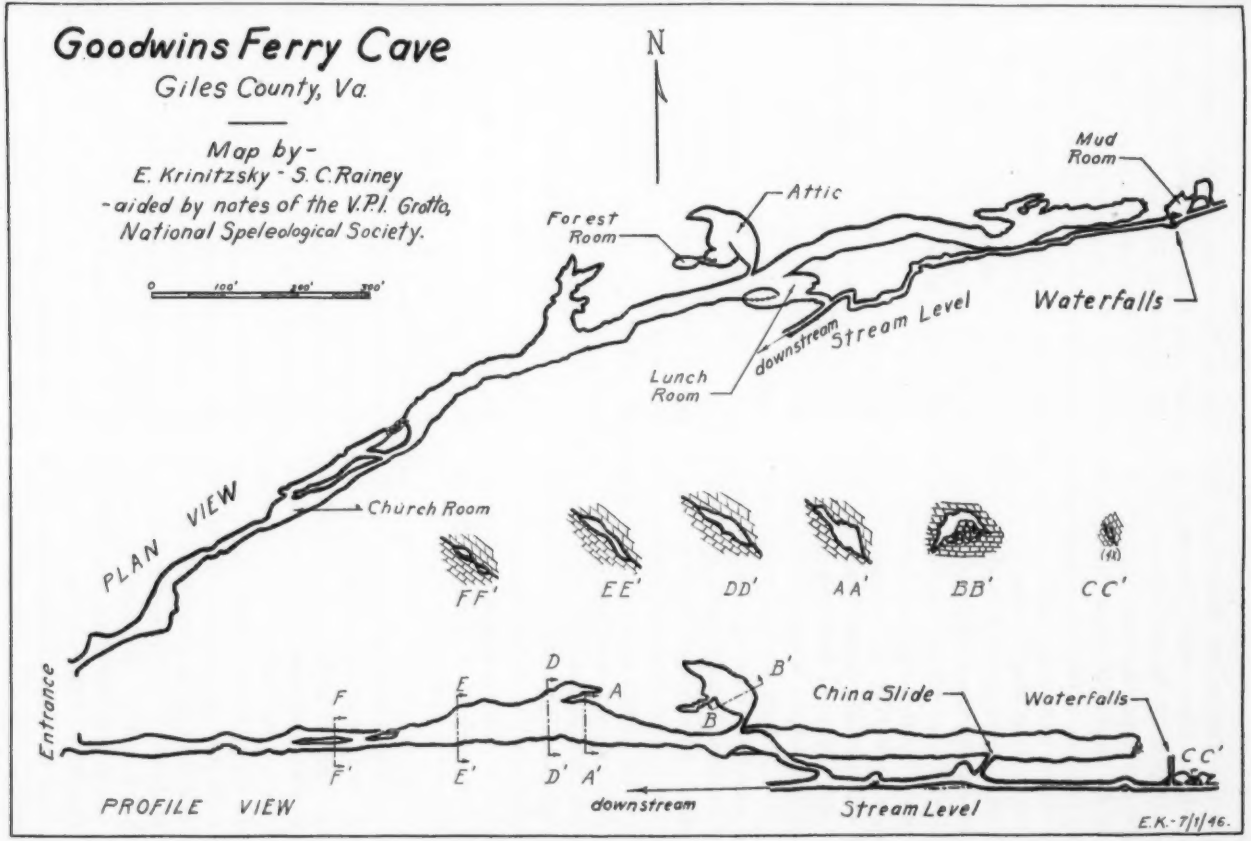

Goodwins Ferry Cave
Giles County, Va.
Map by-
E. Krinitzsky - S. C. Rainey
-aided by notes of the V.P.I. Grotto,
National Speleological Society.
0 100' 200' 300'
N
Mud Room
Attic
Forest Room
Waterfalls
downstream Stream Level
Lunch Room
PLAN VIEW
Church Room
FF'
EE'
DD'
AA'
BB'
CC'
Entrance
F
E
D
A
B'
B
China Slide
Waterfalls
CC'
F'
E'
D'
A'
PROFILE VIEW
downstream
Stream Level
E.K.-7/1/46.

FIG. 2

along the strike of the fault for ½ mile. The fault dip, which averages 53° through most of the cave, dominates the shape of the passageways (Fig. 2). Sections *EE'* and *DD'* show characteristic cross sections. The stream has worked downward along the fault surface to form a tabular-shaped room. In section *AA'* the same type of stream action has been working down the dip, but large rock-falls have caused an enlargement upward. Debris from the rock-falls has been removed by the stream. Section *BB'* shows enlargement in what was probably a zone of weakness due to pronounced jointing. Single blocks, as large as 50 × 50 feet, have fallen from the roof and have built up a huge mound that apparently could not be removed by whatever water circulation there was at the time of the room's development. This room, known as the "Attic," is not the only one that owes its incongruous shape to lines of weakness other than the fault plane. Similar enlargements are the "Mud Room" and the "Waterfalls." In section *CC'* there is shown a striking drop of the stream level in a recent part of the cavern.

Enlargement of the cavern by stream action appears to have been done almost wholly by solution, and, although the cavern has developed along the fault plane where dolomite is in contact with limestone, solution by the stream has caused enlargement chiefly in the limestone. In the front rooms of the cave the roof and right-hand wall (facing inward) are of dolomite, whereas the floor and left-hand wall are of limestone. At the rear of the first room is an accumulation of fallen blocks, which average 6 × 5 × 3 feet, and all of which appear to be of dolomite. In the split of the passageways, near the "Church Room," there are immense numbers of platelike dolomite blocks. They are roughly circular, and many are from 5 to 15 feet in diameter but are only 10 inches to 3 feet in thickness. Farther inward the roof is still of dolomite, but the floor slope is much obscured by flow stone. Wherever bedrock outcrops, however, it is everywhere of limestone. In the Attic the fallen blocks are all of dolomite. Even on the stream level the same association of limestone floor with dolomite roof and dolomite rock-falls can be seen. From the foregoing it may be said that all the rock-falls are probably of dolomite and that enlargement of the cavern into the dolomite has proceeded solely in this manner.

The extremely linear development of the cave (Fig. 2) was caused by its situation on the fault plane, but the horizontal nature of the passageways calls for some other type of control in their development.

II. HYDROLOGY

Shortly after the cave stream ceased flowing in the upper level of the cave, the outlets of the center rooms became dammed, and ponded waters formed good-sized lakes which are now evidenced by flowstone rims that once marked their periphery.

The upper levels of the cave are at present generally dry except for small trickles in the middle rooms which aid dripstone deposition and which form a few tiny pools even now. It may be that at an earlier time there were heavier seepages, the water of which has been diverted or the fissures sealed with dripstone. Prolonged rainy weather moderately increases the number and volume of the trickles and the general dampness of the cave so that there is a greater possibility that the earlier flow of water has been diverted from the fissures.

Dolomite rock-falls are common on the present stream level, but they never have the effect of damming the stream. The water readily percolates through all the rock-fall accumulations that strew the stream's path, and so it can be reasoned that caving of the dolomite roof could not be solely responsible for the damming that took place in the upper level. Most likely the water that seeped into the room was heavily charged with dissolved materials and, by depositing dripstone, served to seal off rock-fall accumulations and so in itself created the barriers to its movement.

At the lowest level there is a permanent stream which flows westward to the New River. In places the stream cross section is roughly $3\frac{1}{2}$ feet wide and $1\frac{1}{2}$ feet deep, but in general the stream flows over a wider, rocky bed. At the Waterfalls (Fig. 2) there is a chimney with a vertical height of 80 feet and a circular cross section around 11 feet in diameter. A heavy fall of water furnishes roughly a third of the volume of the stream.

At one time the stream flowed through the upper level and out of the present cave entrance and there cascaded down the slope as the stream now does at the lower level. There yet remains directly below the entrance a pronounced hollow filled with rocky debris through which the stream originally flowed. At present there is a built-up ledge at the entrance, over which one descends to enter the cavern. No doubt this ledge is the result of the accumulation of rock and soil talus washed down from the hillside above since the stream abandoned this exit.

On the Cave Hill, 75 feet above the road, the cave stream emerges through rock talus as a spring. This emergence is at the level of the stream within the cave, but it accounts for only about half the volume of the cave stream. Below the road level and 70 feet to the south of the cave spring, another spring issues from talus, its volume perhaps somewhat increased by near-surface circulation. It contains the remainder of the cave water. Fluoroscein tests made on the cave waters showed distinct coloration in both springs. The downstream level of the cave can be followed almost to the outlet of the upper spring, but the points at which the water turns down to the lower spring cannot be determined with exactness, as the water is lost gradually.

Pebbles of quartz, limestone, and dolomite occur along the stream bed. Most abundant are those of dolomite, which are sharply angular and are nearly all solution-pitted, especially along lines of joint or fracture weakness. The quartz pebbles are all well rounded and, in all probability, are from the basal conglomerate in the overlying Clinch sandstone and were not rounded by the action of the cave stream. The presence of these quartz pebbles from the Clinch conglomerate indicates a water circulation from the overlying sandstones toward the cave.

The cave stream is remarkably regular in flow, being little affected by rainstorms, and the absence of any sinkhole pattern or major sinkholes on the surface further indicates that the water in the cave must come from the sandstones. But, from the vitreous and compact nature of the sandstones, it seems unlikely that they could act as a porous reservoir. Jointing, however, is fairly well developed in the Clinch sandstone, and seepage along such fissures may be in considerable volume. The synclinal structure of the Juniata shale (Fig. 1), by acting as a basin and keeping a perched water table in the sandstones, could furnish an over-

lying source of water for the cavern. Rainfall in Giles County averages 39 inches per annum and is fairly evenly distributed.[4] Monthly precipitation is around 3.2 inches, with a high of 4.28 inches in July and a low of 2.44 inches in October, the remaining months showing fairly uniform rainfall. It is highly probable that the uniform rainfall along with slow seepage through the sandstones and from the perched water table are responsible for the steady flow of the cave stream.

Fine muds and sands are absent in the stream channel except for small quantities associated with pebble accumulations in hollows. Where the stream beats against fallen blocks of dolomite, there is considerable indication of solution hollowing in the form of pits and grooves on the rocks. Thus it seems likely that abrasion plays but a small part in the stream's erosion and that solution is the principal factor.

III. DEPOSITS

The first two rooms from the entrance are bare except for an occasional nub of a stalagmite, but beyond, on the same level, the rooms have a complete veneer of beautiful flowstone with many elephant's-ear pendants, magnificent stalagmite masses, and several columns. The Lunch Room, the Mud Room, and the stream level in between are of comparatively recent origin and appropriately rugged and bare of dripstone. The stream level, downstream from the Lunch Room, has several spots containing choice deposits, but the best development in the entire case is in the Forest Room.

Although there are many small helictites in the main level, these interesting forms are best developed in places on the ceiling of the Attic. The helictites of the Goodwins Ferry Cave (Fig. 3) are characteristically elongated spines of aragonite which radiate outward in all directions. The spines are generally straight and very slender, although in places the needles of aragonite bunch themselves together to form a stout rod which may be slightly curved. Secondary developments of the aragonite needles radiate outward from growth centers on primary stalks. At many points in the cave are found the common aragonite knobs, but in the Attic there is an unusual form in which the knobs appear to have been pushed out on several stalks.

IV. ORIGIN

Since Davis presented his exhaustive treatment of the two-cycle theory of cave origin,[5] it, and subsequent theories by other authors, has focused attention on the extent to which a cave has been phreatic. Davis believed that most caverns originated below the water table and were completely filled with water during the time of their enlargement; later he postulated a change of conditions which lowered the water table and so drained the caves. Then, he thought, a second cycle began, in which cave enlargement either ceased or proceeded at a much slower rate and during which time dripstone deposition began.

If their very beginning is considered, all caves are phreatic in origin, for the first circulation of water in some incipient fissure that originally determined the cave must have completely filled the tiny course. When a cave is below the water table or in a saturated zone, huge rooms[6]

[4] *Climate and Man: Yearbook of Agriculture* (Washington: U.S. Department of Agriculture, 1941), p. 1159.

[5] W. M. Davis, "Origin of Caves," *Bull. Geol. Soc. Amer.*, Vol. XLI (1930), pp. 475–628.

[6] B. C. Moneymaker, "Subriver Solution Cavities in the Tennessee Valley," *Jour. Geol.*, Vol. XLVIII (1941), pp. 74–86.

can develop from small circulations, and such rooms will be completely filled with water. However, if the cave is not in a water-saturated horizon, the cave void cannot be at all times completely filled by the cave stream unless the volume of the cave is equal to or less than the volume of the stream at its normal rate of flow. When such a stream enlarges its path to a point where the cavern is larger than the volume of the stream, the cave ceases to be phreatic. But enlargement should not slow down at this time. As long as the stream does not leave the enlarged cave course, and as long as the rate of circulation does not diminish, the stream will continue to excavate as it did previously, and, at the same time, dripstone may be deposited. The author believes that such conditions prevailed in the development of at least the lower level in the Goodwins Ferry Cave.

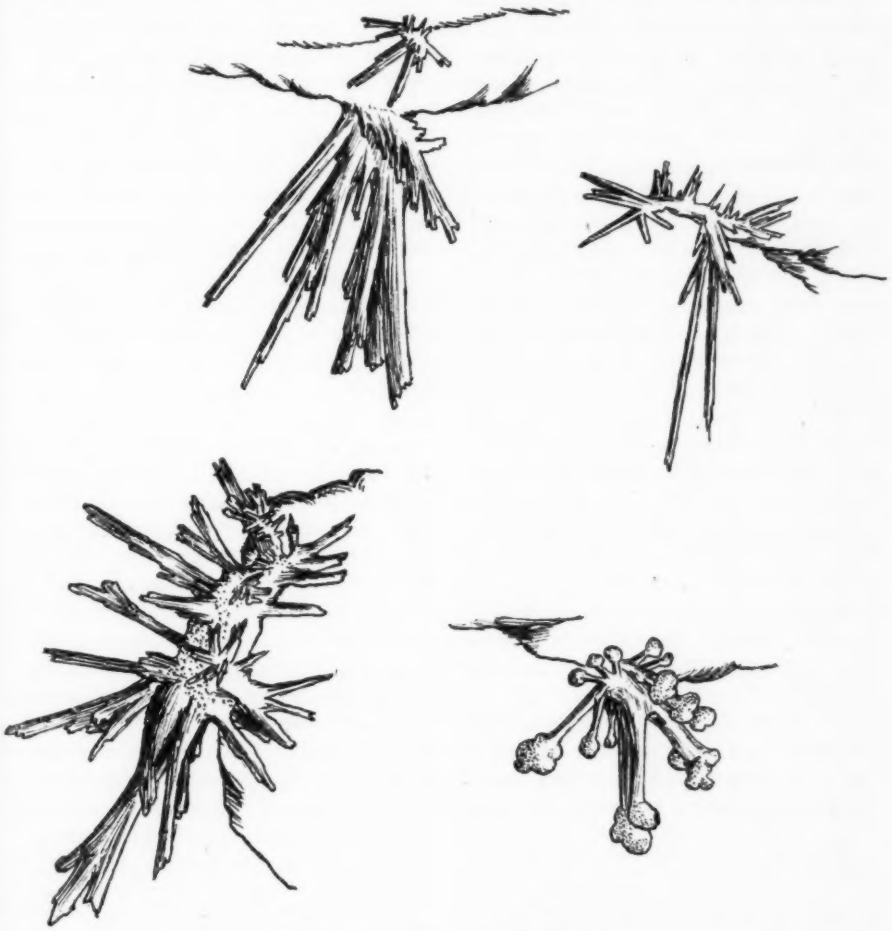


FIG. 3.—Typical helictites (natural size)

Malott[7] postulated another aspect of the Davis theory by asserting that cave systems develop in a spongelike or rambling pattern below the water table. Later these passages are drained, when elevated above the water table, and are then invaded by surface streams which work into the pre-existing cave system and modify the caves according to the more linear nature of their erosion. The Goodwins Ferry Cave is unusually straight and level with no evidence of earlier rambling channels. Further, there is no evidence, on the slope of Spruce Run Mountain, of the remains of any surface stream which could have been located so as to work its way into the cave. As has been indicated, the source of the cave's water is, and probably always has been, from the overlying sandstones. Surface streams are frequently subject to floods, and underground streams of surface origin are also subject to fluctuations in volume due to intermittent rainfall. The lower level of the Goodwins Ferry Cave does not show any evidence of flooding. The ceiling, which in many places slopes down to the stream, is entirely clean and shows no mark of higher water levels. Further, the cave water is clear and, to the author's knowledge, has never been known to become muddy, as would be expected if the water came from a surface stream. It seems certain that the invasion theory in its original form does not apply to the Goodwins Ferry Cave.

Still along a similar line, Swinnerton[8] presented the theory that lateral flow in the upper zone of the water table causes the development of caverns. He, also, was trying to explain originally rambling patterns in caves as well as the horizontal nature of many cave levels. In a limited sense, this theory could apply to the Goodwins Ferry Cave, since the stream in old age would be flowing on the level of a water table. But, since there are no rambling patterns, and the stream's water comes from the overlying sandstones, the lateral-flow theory in its proper form is not applicable to this cave.

The concept of an overlying water table that furnishes the source of a cave stream was advocated by Gardner,[9] who, while admitting other modes of origin, stressed the theory that caves develop from the tapping of static-water zones as surface valleys are deepened in the process of baseleveling. Undoubtedly, the Goodwins Ferry Cave furnishes a path for the drainage of water from the sandstones that make up the top of Spruce Run Mountain down to the New River which determines the local baselevel.

The New River, in cutting across Spruce Run Mountain, has left a steep gorge, on one side of which the entrance to the cave is perched. The extreme levelness of the cave (Fig. 2) indicates that it must have developed when the New River was flowing near the elevation of the cave, which is now about 250 feet above the river level. It has been noted that the cave stream is in the act of lowering itself to the present course of the New River through the lower spring, which drains about half of the cave water and which issues at practically the same elevation as the river. It seems logical to suppose that the New River was at one time stationary at or near the elevation of the cave entrance long enough to furnish time for the cave to develop strongly on

[7] C. A. Malott, "Invasion Theory of Cavern Development," *Proc. Geol. Soc. Amer. 1937*, p. 323.

[8] A. C. Swinnerton, "Origin of Limestone Caverns," *Bull. Geol. Soc. Amer.*, Vol. XLIII (1932), pp. 663–94.

[9] J. H. Gardner, "Origin and Development of Limestone Caverns," *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), pp. 1255–74.

that horizon and that, since that time, the river has been downcutting more rapidly than the cave stream. However, the author has not been able to recognize any surface indications of a pause in the river's erosion or any terraces or other erosional surfaces near by at elevations that might suggest a geomorphic relation to the cave. It is possible that the cave may be a more delicate criterion of temporary regional baselevel than are surface erosional features. If this is the case, the study of other caves in the region could furnish comparable data on elevations which might prove to be of significant value.

The volume of the cave stream, when it was active in the upper level of the cave, cannot be determined, but, from the present exceedingly steady flow and the source of the water, there is no reason to think that the stream's volume has varied much, barring climatic changes. The hollow through which the stream cascaded when flowing out of the upper level is strikingly similar to the little valley of the present stream farther down the slope, both in size and in talus accumulation. Thus it seems likely that there was a similarity in stream size throughout the time that the cave has existed.

Inside the cave the huge development of the Attic and the other similar enlargement about 150 feet west (Fig. 2) seem incongruous when compared with the remainder of the cave. Excessive rock-falls in the Attic have obscured much of the true nature of that room, but it is evident that much stream erosion took place there at an earlier time. It is highly probable, and in keeping with the rest of the cave, that the cave stream at one time entered the Attic in some manner before the main passage had worked its way beyond the Attic and that the stream flowed from the present tip of the Attic directly across to the previously mentioned enlargement and so developed it. The extension of the Forest Room is probably indicative of another channel, in the same direction, but developed at a later time. Rock-falls and dripstone deposition could easily have sealed off such connections and also obscured evidences of stream entrance into the Attic. Since there are about 200 feet of limestone between the Attic and the source of the stream in the overlying sandstone, it is quite likely that there once existed many ramifications of the cavern that have been lost. The tiny outlet of the Attic into the present hallway indicates that the stream flowed through that opening for only a comparatively short time before it left the Attic entirely and entered the main hallway somewhere beyond its present extremity, which is now blocked by rock-falls. A few abandoned and blocked courses seem to have developed about 200 feet west of the present termination of the main hallway. Further confirmation of the flow from the tip of the Attic across to the hall enlargement is afforded by the excessive height of the main hall west of this enlargement. The height not only is due to flow from two sources but also may indicate a strong slope in the first channel, as there does not seem to be any excessive widening farther west toward the entrance. The development and abandonment of several stream courses in the main level are also shown directly east of the Church Room.

After the mature development of the upper level came the ponding already referred to. For a while it is likely that the stream flowed in the upper level to the Lunch Room, developing that room to its present size, and at the Lunch Room the stream then took the course down to the

lower level. Next in the development of the cave was the channel called the "China Slide," which pirated the stream's water away from the course leading to the Lunch Room and increased the development of the cave's lower level. Following the China Slide cutoff, part of the stream appears to have found access to the lower cave at a distance of more than 300 feet upstream. Here, two moderately sized rooms were developed, and a waterfall of recent origin seems to be tapping a stream in what may be an extension of the upper hallway blocked by debris. A difference in age can be seen on the lower cave level, as there is dripstone downstream from the Lunch Room, but the stream level is devoid of deposits for its entire length upstream. It has already been stated that the stream at present loses much of its water at points downstream from the Lunch Room. Evidently the development of the cave takes place in regularly repetitive steps in which the stream continuously works downward and, once achieving a lower level, backward to regain any length it may have lost.

The waterfall is an extremely youthful feature, judging from its diameter of only about 11 feet. The volume of water crashing down the center of the chimney produces a spray that keeps the cylindrical walls constantly wet, and it is probably this descending film of water that dissolves the wall rock and thus enlarges the chimney. As the chimney becomes enlarged most rapidly downward, the water goes over a lip at the top to fall freely in the center of the chimney. Further development would result in a dome room.[10] Such waterfall tapping of an overhead stream is called "subterranean stream piracy."[11] In this instance water has been diverted from another course into that of a more youthful and more vigorous stream.

The caving of the dolomite roof was an important factor in enlarging the cavern, but the ceiling everywhere appears extremely firm, and it is not apparent why such large rock-falls should have taken place. Of course, enlargement of the cavern may have enabled rock pressures, together with original open fissures in the dolomite, to drop the blocks. But there is a possible tectonic explanation, for the center of a recent major earthquake was only 10 miles from the cave. Following are some excerpts from a report on the shocks:

> Immediately following the earthquake of May 31, 1897, which was distinctly felt over most of the eastern portion of the United States, came reports of continued disturbances in the form of explosions and earth tremors in Giles County, Virginia. . . . . In Pearisburg . . . . no serious damage was done but old brick houses were badly shaken, and many chimneys were cracked and the topmost bricks hurled to the ground. Much noise accompanied the shock. . . . . At Narrows . . . . the surface is said to have rolled like the groundswells of the ocean, springs were muddied and in some cases ceased to flow for a short time after the shock occurred, and a landslide of considerable proportions and a big rock rolled down off the face of Wolf Creek Mountain.[12]

No doubt such shocks could have caused rock-falls in the cave, but whether or not most of the rock-falls depended upon such unusual disturbances is an unanswerable question. Gardner[13] believed that unrecorded earthquakes were important in causing enlargements by caving in Carlsbad Cavern and other west-

[10] J H. Bretz, "Vadose and Phreatic Features of Limestone Caverns," *Jour. Geol.*, Vol. L (1942), p. 683.

[11] *Ibid.*, p. 685.

[12] M. R. Campbell, "Earthquake Shocks in Giles County, Virginia," *Science*, Vol. VII (1898), pp. 233–35.

[13] Pp. 1272 of ftn. 9.

ern caves, but these opinions also cannot be proved.

V. OTHER FAULT CAVES

In the American literature the writer has found no similar caverns described, and, for the present, the Goodwins Ferry Cave and its sister-cavern across the river may be considered as being unique in their association with a major fault.

It has not been possible to investigate the foreign literature thoroughly, owing to the lack of comprehensive foreign bibliographies prior to 1933. The only reference to a similar cave that the writer has been able to find is the report of C. Sorotchinsky on the origin of the Salle du Dôme in the Grotte de Han of Belgium.[14] The origin of the "Dome Room" is attributed to a combination of faulting, caving, and solution, but it differs from the Goodwins Ferry Cave in that the room does not lie on the fault itself but is on one side of a faulted zone. Enlargement of the room has resulted from caving of the ceiling in a limestone bent into a monocline. The Dome Room has a height of over 100 feet and is located under the crest of the monocline which made the enlargement possible. The room is not associated with the fault, but near by is a stream in a constricted passageway ("passage du Diable") which appears to be well in the fault zone. The fault cannot be defined as a plane but is represented by a zone of brecciation.

[14] C. Sorotchinsky, "Un accident téctonic éclairant la génèse de la Salle du Dôme dans la Grotte de Han," *Bull. Soc. Sci. Bruxelles*, Vol. LIX (1939), pp. 97–106.

CONCLUSIONS

The Goodwins Ferry Cave is the result of solution along the plane of weakness of the Saltville fault. The cave stream has worked its way down the dip of the fault and along the fault strike, chiefly in the limestone. Rock-falls from the roof have enlarged the cavern into the dolomite, and, as the cavern works its way down the dip, dripstone is acting to fill in the cavern from above. The linear shape of the cave is due to its location on the fault plane, and the strongly horizontal development may be due to an earlier control of the cave's drainage by the New River.

ACKNOWLEDGMENTS.—The author is indebted to Dr. J. W. Huddle, of the University of North Carolina, for much valuable assistance in the preparation of this paper. Dr. B. N. Cooper, of Virginia Polytechnic Institute, has given advice on the field work; and field assistance was received from Messrs. S. C. Rainey and I. R. Tannenbaum of the Pioneer's and Hobo's Union. The writer also deeply appreciates the hospitality of Mr. A. B. Scott of Goodwins Ferry, Virginia.

REVIEWS

Geology of Florida. By C. WYTHE COOKE. (Florida Geological Survey Bull. 29.) Prepared in co-operation with U.S. Geological Survey. Tallahassee, Fla., 1945. Pp. ix+339; pls. 1 (map in pocket, scale 1/1,000,000); figs. 47.

It has been sixteen years since Cooke and Mossom's *Geology of Florida*, published as *Florida Geological Survey Twentieth Annual Report*, first appeared. During that time a large amount of new information has been gathered dealing with the geology of Florida; and, although much of it has been published elsewhere, there was great need for a new bulletin bringing this information together in one place and interpreting it in the light of the latest knowledge.

Dr. Cooke has done this in Bulletin 29, which bears the same title as his and Mossom's earlier report.

The new bulletin contains much that is recognizable in the older report with only such references to it as Cooke believed to be of historical interest. As he remarks in the Preface, "It contains many paraphrases and verbatim excerpts from that report without quotation marks or other indications of origin." Thus, Bulletin 29 is more a revision of an old, standard work than an entirely new treatise. It is essentially a stratigraphic paper; however, its contents are not limited to stratigraphy. Five pages are given to a lucid description of the composition and structure of the Floridian Plateau, six pages are devoted to topography, and four pages describe the effect of an oscillating sea-level during Pleistocene time. The Bibliography requires eleven pages and the Index twelve; the balance, two hundred and ninety-seven pages, describes the stratigraphy.

Cooke describes each formation according to the following plan: A, general features: (1) name, (2) characters, (3) thickness, (4) distribution, (5) stratigraphic relations, (6) paleogeography, (7) fauna, and, occasionally, (8) utilization; and B, local details (which includes formation descriptions by counties). In general, the plan and treatment of the subject are excellent.

Since the writing of the *Twentieth Annual Report* a considerable amount of exploration of the subsurface formations by means of test-well drilling has been done by oil companies and independent operators over much of the state in a search for petroleum and by the U.S. Geological Survey in southern Florida on ground-water research. The data uncovered have led to recognition of several new stratigraphic units and, together with new discoveries of surficial relationships, have aided greatly in the solving of several problems in stratigraphy and in the geological history of the state. These new findings are ably discussed by Cooke.

Owing to a lack of printing funds, the geological map of the state (Pl. I) as a whole does not compare favorably with the earlier geological map published to accompany the *Twentieth Annual Report*. The scale would have been more acceptable, especially to petroleum geologists, had it been the same as that used by the state geological surveys of Alabama and Georgia. Use of colors in patterns of blue and orange is not as satisfactory as the color scheme used on the old map, nor is the means of showing the distribution of Pleistocene formations entirely suitable. For example, the Anastasia formation, which is largely contemporaneous with the Miami oölite and Key Largo limestone, is included with the higher terrace deposits under the label "Late Pleistocene Deposits." To be consistent, the Miami oölite and Key Largo limestone also should have been included. In addition, there are vast areas of the state that are covered with a mantle of Late Pleistocene materials; these are now shown but could have been as easily as the areas delineated. It would seem to have been better either to have eliminated entirely the higher Pleistocene terrace deposits (as Cooke did for much of the state) or to have shown their presence over the underlying formations by use of a distinctive overprint.

The route of part of the proposed Florida ship canal is indicated prominently on the map; however, the canal is not yet constructed nor is there any assurance that it will be built. The route of the proposed canal as shown on the map may also lead one to wonder if the eastern terminus is to be at St. Augustine or at the mouth of the St. John's River.

In general, Bulletin 29 is an excellent summary of the stratigraphy of Florida as shown at the time of writing. It fills a need felt by professional geologists, amateurs, and laymen since the *Twentieth Annual Report* went out of print several years ago.

GARALD G. PARKER

PUBLICATIONS RECEIVED

The Age, Origin, and Classification of the Rivers of India. By H. L. Chhibber. National Geographical Society of India Bulletin 1. Benares, 1946.

The Agricultural Development of the Middle East. By B. A. Keen. A Report to the Director General, Middle East Supply Centre. London: H.M. Stationery Office, May, 1945.

Agstone Used in Illinois in 1945. By Walter H. Voskuil and Douglas F. Stevens. Illinois State Geological Survey, Circular 123. Urbana, July, 1946.

Bibliography of the Geology and Natural Resources of North Dakota. By Chrissie E. Budge. North Dakota Research Foundation Bulletin 1. Bismarck, 1946.

Biokronologisk Datering vid en Naturpegel, Penningby Gamla Slott. By Ebba Hult de Geer. Fr. Stockholms Högskolas Geokronol. Inst., Geol. Fören. Förhandl., Vol. LXXIV, 1945.

Bulletin de la commission geologique de Finlande. No. 136. Containing various papers. Suomen Geol. Toimikunta. Helsinki, December, 1945.

A Census of the Determinable Genera of the Stegocephalia. By E. C. Case. Transactions of the American Philosophical Society, New ser., Vol. XXXV, Part IV. Independence Square, Philadelphia 6: American Philosophical Society, September, 1946.

Census of Western Australian Minerals. By Dorothy Carroll. Department of Mines, Mineral Resources of Western Australia Bulletin 1. Perth: Robert H. Miller, Government Printer, 1945.

Contributions to the Paleontology of the Lebanon Mountains, Republic of Lebanon. Part 3. The Pelecypod Fauna of the "Olive Locality" (Aptian) at Abeih. By Harold E. Vokes. Bulletin of the American Museum of Natural History, Volume LXXXVII, Art. 3. New York, 1946.

Cretaceous Stratigraphy of the Belvidere Area, Kiowa County, Kansas. By Bruce F. Latta. University of Kansas Publications, State Geological Survey of Kansas Bulletin 64, Part 6. Lawrence, 1946.

The Diptera or True Flies of Connecticut, Second Fascicle, Family Culicidae, the Mosquitoes. By Robert Matheson. State of Connecticut, Geological and Natural History Survey Bulletin 68. Hartford, 1945.

Directory of Georgia Mineral Producers 1946. Georgia State Division of Conservation, Geological Survey, Department of Mines, Mining and Geology. Atlanta, July, 1946.

Les Doctrines de Georges Cuvier dans leurs rapports avec le transformisme. By Émile-Georges Dehaut. Encyclopédie Biologique, Vol. XXIV. Paris: Paul Lechevalier, 12, Rue de Tournon, 1945.

Economic Geography of Canada. By A. W. Currie. Toronto: Macmillan Co. of Canada Ltd., 1945. $3.50.

Estudos brasileiros de geologia: Estudo morfológico da Bartina de Camamú. By Elysiario Tavora Filho. Fundação Getúlio Vargas, Fasc. 2, Ano 1, Vol. 1, March, 1946. Rio de Janeiro: Livraria Kosmos, 1946.

Estudos brasileiros de geologia: Ocorrências de calcário no Rio Grande do Sul e prospeção da jazida de "Vacacai" no município de São Gãbriel. By Viktor Leinz. Fundação Getúlio Vargas, Fasc. 1., Ano 1, Vol. 1 (February, 1946). Rio de Janeiro: Livraria Kosmos, 1946.

Exploration for Oil and Gas in Western Kansas during 1945. By Walter A. Ver Wiebe. University of Kansas Publications, State Geological Survey of Kansas Bulletin 62. Lawrence, June, 1946.

Fifth Biennial Report of the State Department of Geology of the State of Oregon, July 1, 1944 to July 1, 1946. Department of Geology and Mining Industries Bulletin 32. Portland, 1946.

Fluorspar Deposits of the Jamestown District, Boulder County, Colorado. By E. N. Goddard. Proceedings of the Colorado Scientific Society, Vol. XV, No. 1. Denver, 1946.

Fluorspar Resources of New Mexico. By Howard E. Rothrock, C. H. Johnson, and A. D. Hahn. New Mexico Bureau of Mines and Mineral Resources Bulletin 21. Socorro, N.M., 1946.

Formulaire technique du préhistorien: Ce qu'il faut savoir sur la géologie et la biogéographie du Quaternaire, l'archéologie et l'anthropologie préhistoriques. By Raymond Furon. Paris: Paul Lechevalier, 12, Rue de Tournon, 1945.

Fossil Penguins. By George Gaylord Simpson.

Bulletin of the American Museum of Natural History, Vol. LXXXVII, Art. 1. New York, 1946.

The Geologic History of the Bison in the Great Plains (A Preliminary Report). By C. Bertrand Schultz and W. D. Frankforter. Bulletin of the University of Nebraska State Museum, Contribution of the Division of Paleontology. December, 1946.

Geological Aspects of Prospecting and Areas for Prospecting in the Zinc-Lead District of Northwestern Illinois. By H. B. Willman, R. R. Reynolds, and Paul Herbert, Jr. Illinois Geological Survey, Report of Investigations No. 116. Urbana, 1946.

The Geological History of Queensland: A Stratigraphical Outline. By W. H. Bryan and O. A. Jones. University of Queensland Papers, Department of Geology, New ser., Vol. II, No. 12, 1945. Published April, 1946. *7s. 6d.*

Geological Work on the Great Artesian Basin. By F. W. Whitehouse. University of Queensland Papers, Department of Geology, New ser., No. 18. Brisbane: A. H. Tucker, Government Printer, 1945.

Geology and Coal Occurrences of Wapiti-Cutbank Area Alberta. By J. A. Allan and J. L. Carr. Province of Alberta, Research Council of Alberta, University of Alberta Report No. 48. Edmonton, 1946.

Geology of the Coastal Plain of East-Central Georgia. By Philip E. La Moreaux. Georgia State Division of Conservation, Department of Mines, Mining and Geology. Geological Survey Bulletin 50, Part 1. Atlanta, 1946.

Geology and Ground-Water Resources of the Coastal Plain of East-Central Georgia. By Philip E. La Moreaux. Georgia State Division of Conservation, Department of Mines, Mining and Geology. Geological Survey Bulletin 52. (Published in co-operation with the Geological Survey, U.S. Department of Interior, Washington, D.C.) Atlanta, 1946.

Geology and Ground-Water Resources of Grant, Haskell, and Stevens Counties Kansas. By Thad G. McLaughlin. University of Kansas Publications, State Geological Survey of Kansas Bulletin 61. Lawrence, July, 1946.

The Geology of the Mt. Washington Quadrangle, New Hampshire. By Marland P. Billings, *et al.* New Hampshire Planning and Development Commission. Concord, 1946.

Geology of the St. Helens Quadrangle, Oregon. By W. D. Wilkinson, W. D. Lowry, and E. M. Baldwin. State of Oregon, Department of Geology and Mineral Industries, Bulletin 31. 702 Woodlark Bldg., Portland 5, Ore., 1946.

Les Gisements de terres plastiques et réfractaires d'Andenne et du Condroz. By Leon Calembert. Université de Liége, Institut de Géologie. H. Vaillant-Carmanne, S.A., Imp. de l'Académie. Liége, 1945.

Glimpses from Resource-full Kansas. University of Kansas Publications, State Geological Survey of Kansas. Lawrence, 1946.

Ground-Water Conditions in Arkansas River Valley in the Vicinity of Hutchinson, Kansas. By Charles C. Williams. University of Kansas Publications, State Geological Survey of Kansas Bulletin 64, Part 5. Lawrence, 1946.

Ground-Water Conditions in Elm Creek Valley, Barber County, Kansas: With Special Reference to Contamination of Ground Water by Oil-field Brine. By Charles C. Williams and Charles K. Bayne. University of Kansas Publications, State Geological Survey of Kansas Bulletin 64, Part 3. Lawrence, 1946.

The Ground-Water Resources of the Glacial Deposits in the Vicinity of Canton, Ohio. By Edward J. Schaefer, George W. White, and Donald W. Van Tuyl. Ohio Water Resources Board, Columbus, Ohio, in co-operation with the U.S.G.S. and the City of Canton, Ohio, Bulletin 3. June, 1946.

Hollidaysburg-Huntingdon Folio, Pennsylvania. By Charles Buts. Geologic Atlas of the United States. Washington, D.C.: U.S. Department of Interior, Geological Survey, W. E. Wrather, Director, 1945.

The Illinois State Geological Survey in War Mineral Research. By M. M. Leighton. Illinois State Geological Survey, Circular No. 121. Urbana, 1945.

India, Part I: Physical Basis of Geography of India. By H. L. Chhibber. Benares: Nand Kishore & Bros., 1945.

Intraspecific Variation in, and Ontogeny of, *Prionotropis woollgari* and *Prionocyclus wyomingensis*. By Otto Haas. Bulletin of the American Museum of Natural History, Vol. LXXXVI, Art. 4. New York, 1946.

The Iron Ore Bearing Formations of Ohio. By Wilber Stout. Geological Survey of Ohio, 4th ser., Bulletin 45. Columbus, 1944.

Laboratory Exercises in Physical Geology. By William C. Putnam and Robert W. Webb. 2d ed. Stanford University, Calif.: Stanford University Press, 1945.

Lee County Mineral Resources. By Franklin Earl Vestal. Mississippi State Geological Survey Bulletin 63. University, Miss., 1946.

Mica Deposits of the Petada District Rio Arriba County, New Mexico. By Richard H. Jahns. New Mexico Bureau of Mines and Mineral Resources Bulletin 25. Socorro, N.M., 1946.

Middle East Science. By E. B. Worthington. A Report to the Director General, Middle East Supply Centre, August, 1945. London: H.M. Stationery Office, 1946.

The Mineral Resources of Western Australia. Prepared in the Government Mineralogical and Chemical Laboratories. 5th ed. Perth, Western Australia, 1945.

A New Anosteirine Turtle from Manchuria. By Ranier Zangerl. Fieldiana—Geology. Chicago Natural History Museum, Vol. X, No. 3. January 23, 1947.

New Mexico Bureau of Mines and Mineral Resources, Annual Report 1, for the fiscal year July 1, 1945—June 30, 1946. By E. C. Anderson, Director. Socorro, N.M., 1946.

The Occurrence of Flint in Ohio. By Wilber Stout and R. A. Schoenlaub. Geological Survey of Ohio. 4th ser., Bulletin 46. Columbus, 1945.

Oil and Gas Development in Illinois in 1945. By Alfred H. Bell and Virginia Kline. Illinois Geological Survey Press Bulletin Series, Illinois Petroleum, No. 54. Urbana, 1946.

Om Mineraljordarternas Hållfasthetsegens ("On the Strength of Properties of Mineral Soils"). By Thord Brenner. Geologinen Tutkimuslaitos. Bulletin de la Commission Géologique de Finlande, No. 139. Helsinki, March, 1946.

On the Geochemical Differentiation in the Earth's Crust. By Kalervo Rankama. Suomen Geol. Toimikunta. Bulletin de la Commission Géologique de Finlande, N. 137. Helsinki, January, 1946.

Perlite Deposits Near the Deschutes River Southern Wasco County, Oregon. By John Eliot Allen. State of Oregon Department of Geology and Mineral Industries, G.M.I. Short Paper No. 16. Portland, 1946. $0.15.

The Petrography of Some Queensland Oil Shales. By A. W. Beasley. Reprinted from Vol. XII, Part III, of the Queensland Museum, August 6, 1945. New ser., No. 20, University of Queensland Department of Geology, 1945.

The Physiography and Igneous Geology of Hongkong and the New Territories. By M. Y. Williams, R. W. Brock, S. J. Schofield, and T. C. Phemister. From the Transactions of the Royal Society of Canada, 3d ser., Sec. IV, Vol. XXXIX, 1945. Ottawa, 1945.

Preglacial Erosion Surfaces in Illinois. By Leland Horberg. Illinois Geological Survey, Report of Investigations No. 118. Urbana, 1946.

Preliminary Report on the Kokomo Mining District, Colorado. By A. H. Koschmann and F. G. Wells. Colorado Scientific Society Proceedings, Vol. XV, No. 2. Denver, 1946.

Proceedings of the Geol. Society of South Africa: Vol. XLV, Containing the Minutes of Meetings and the Discussions on Papers Read during 1942; Vol. XLIV, Containing the Minutes of Meetings and the Discussions on Papers Read during 1941. Edited by the Honorary Secretary. Johannesburg: Hortors, Ltd., Harrison St., 1942, 1943.

Publications on the Geology and Mineral Resources of Virginia. By Arthur Bevan. Virginia Geological Survey, Virginia Conservation Commission, Circular 2, Suppl. 1. University, Va., July, 1946.

Reconnaissance Geology of Limestone Deposits in the Willamette Valley, Oregon. By John Eliot Allen. State of Oregon, Department of Geology and Mineral Industries, G.M.I. Short Paper No. 15. Portland, 1946. $0.15.

Report of the Geological Survey for the Year 1944, Western Australia. Perth: By Authority, Robert H. Miller, Government Printer, 1945.

Report of the Government Geologist for the Year 1943. Western Australia Department of Mines. Perth: By Authority, Robert H. Miller, Government Printer, 1944.

Roadstone: Geological Aspects and Physical Tests. By J. Phemister, *et al.* Road Research Special Report No. 3. Department of Scientific and Industrial Research. London: H.M. Stationery Office, 1946.

Sebecus, Representative of a Peculiar Suborder of Fossil Crocodilia from Patagonia. By Edwin Harris Colbert. Bulletin of the American Museum of Natural History, Vol. LXXXVII, Art. 4. New York, 1946.

Silicified Rock in the Ogallala Formation. By John C. Frye and Ada Swineford. University of Kansas Publications, State Geological

Survey of Kansas Bulletin 64, Part 2. Lawrence, 1946.

Sodium Salts of Lake County, Oregon. By Ira S. Allison and Ralph S. Mason. State of Oregon, Department of Geology and Mineral Industries, G.M.I. Short Paper No. 17. Portland, 1947.

Some Notes on the Archaeology of the Department of Puno, Peru. By Marion H. Tschopik. Expeditions to Southern Peru, Peabody Museum, Harvard University, Report No. 3. Papers of the Peabody Museum of American Archaeology and Ethnology, Harvard University, Vol. XXVII, No. 3., 1946.

Tantalum and Niobium. By Keith R. Miles, Dorothy Carroll, and H. P. Rowledge. Department of Mines, Mineral Resources of Western Australia. Perth: By Authority, Robert H. Miller, Government Printer, 1945.

Texas Mineral Resources. University of Texas Publication No. 4301. Bureau of Economic Geology. Austin, Tex., November, 1946.

Transactions of the Geol. Society of South Africa, Vols. XLIV, XLV, XLVI, XLVII, and XLVIII, Containing Papers Read during 1941, 1942, 1943, 1944, and 1945. Edited by the Honorary Secretary. Johannesburg: Hortors, Ltd., Harrison St., 1942, 1943, 1944. 1945, and 1946.

Use of Electrical Geophysical Methods in Groundwater Supply. By Carl A. Bays. Illinois Geological Survey, Circular 122. Urbana, 1946.

Well-Sample Records. By James H. C. Mattens. West Virginia Geological Survey, Vol. XVII, 1945.

What about Our Minerals? Illinois Geological Survey, Circular 124. Urbana, 1946.

MAPS RECEIVED

Bathymetrical Map in Seven Sheets of the Norwegian Coastal Waters and Adjoining Seas. Map accompanying Olaf Holtedahl, "The Submarine Relief off the Norwegian Coast." Oslo: Norske Videnskaps-Akademi, 1940.

Geologic Map and Structure Sections of the Mascot–Jefferson City Zinc Mining District, Tennessee. By Josiah Bridge. Scale 2 inches to 1 mile, 1/31,690. State of Tennessee, Department of Conservation, Division of Geology, 1945.

Geological Map of South America, Part 2 (Preliminary). Geological Society of America, Foreword and explanation legend in English, Portuguese, and Spanish by A. I. Levorsen. Scale 1/5,000,000. Washington, D.C.: Williams & Heinz, 1945.

Map of Barnsley, Ordnance Survey of England, Sheet 87. Scale 1 inch to 1 statute mile, 1/63,360. 3d ed. Geological Survey of England and Wales, E. B. Bailey, Director, 1942.

Mineral Locality Map of Texas. By E. H. Sellards, Glen L. Evans, and Leo Hendricks. University of Texas Publication No. 4301, Plate I. Scale 1/1,000,000. Bureau of Economic Geology. Austin, Tex., November, 1944.

Physiographic Diagram of Africa. By A. K. Lobeck. Scale 1/14,000,000. New York: Geographical Press, Columbia University, 1946.

State of Oregon, Map Showing Location of Quicksilver Deposits. Prepared by Francis Frederick, 1945. Scale 1/1,000,000.

State of Oregon, Map Showing Principal Mineral Deposits. Scale 1/1,000,000. Department of Geology and Mineral Industries, F. W. Libbey, Director, 1946.